LIBRARY
FISH & WILDLIFE SERVICE
PACIFIC OCEANIC FISHERY
INVESTIGATIONS

SEARS FOUNDATION FOR MARINE RESEARCH
BINGHAM OCEANOGRAPHIC LABORATORY, YALE UNIVERSITY

JOURNAL OF MARINE RESEARCH

VOLUME VII, 1948

NUMBER 2

ISSUED SEPTEMBER 15, 1948, NEW HAVEN, CONN., U. S. A.

JOURNAL OF MARINE RESEARCH

PUBLISHED BY

SEARS FOUNDATION FOR MARINE RESEARCH
BINGHAM OCEANOGRAPHIC LABORATORY, YALE UNIVERSITY

BOARD OF EDITORS

Chairman: A. E. PARR
American Museum of Natural History, New York

C. O'D. ISELIN
Woods Hole Oceanographic Institution

H. U. SVERDRUP
Scripps Institution of Oceanography

DANIEL MERRIMAN
Bingham Oceanographic Laboratory

C.-G. ROSSBY
University of Chicago

THOMAS G. THOMPSON
Oceanographic Laboratories, University of Washington

MANAGING EDITOR

YNGVE H. OLSEN
Bingham Oceanographic Laboratory
Yale University
New Haven, Conn.

The Journal is normally issued in three numbers per year constituting one volume, except for volume V, published in 1942–1944, and volume VI, published in 1946 and 1947. All the income derived from subscriptions will be used solely to increase the volume and improve the quality of the Journal for the benefit of its subscribers and contributors.

Subscription price three dollars per volume, or one dollar per number. Address all communications to the Managing Editor. See inside rear cover for Note to Authors.

SEARS FOUNDATION FOR MARINE RESEARCH
BINGHAM OCEANOGRAPHIC LABORATORY, YALE UNIVERSITY

JOURNAL OF
MARINE RESEARCH

VOLUME VII 1948 NUMBER 2

THE MEASUREMENT OF THE VISCOSITY COEFFICIENT OF SEA WATER

By

YASUO MIYAKE and MASAMI KOIZUMI

Central Meteorological Observatory, Japan

In an old study by Ruppin and Krümmel (Krümmel, 1907: 281), the viscosity of sea water was measured by the method of Wil. Ostwald. However, at that time the data for the viscosity of sea water were given relative to that of pure water at 0° C. Since then several precise measurements (Dorsey, 1940: 182) have been made and reliable data obtained, and therefore it is necessary to determine more exact values for the viscosity of sea water.

Method of Measurements. The Ostwald viscometer was used. The size of the capillary was so designed that about 5 cc. of sea water could run through it in five to ten minutes. The temperature variation of a thermostat in which the viscometer was submerged was adjusted to within 0.01° C.

The measurements were made at temperatures near 0°, 5°, 10°, 15°, 20°, 25° and 30° C. Sea water with a chlorinity of 19.39 °/₀₀ was diluted to concentrations which were respectively 2, 4, 6, 8, 10, 12, 14, 16 and 18 °/₀₀; these waters were used as samples.

The coefficients of viscosity can be calculated by the formula

$$\eta = \eta_w \frac{\rho \cdot t}{\rho_w \cdot t_w},$$

where η is the coefficient of viscosity of sea water at any given temperature, η_w the coefficient of pure water at the same temperature, ρ and ρ_w respectively the density of sea and pure water at the given

JOURNAL OF MARINE RESEARCH

PUBLISHED BY

SEARS FOUNDATION FOR MARINE RESEARCH
BINGHAM OCEANOGRAPHIC LABORATORY, YALE UNIVERSITY

BOARD OF EDITORS

Chairman: A. E. PARR
American Museum of Natural History
New York

C. O'D. ISELIN
Woods Hole Oceanographic Institution

H. U. SVERDRUP
Scripps Institution of Oceanography

DANIEL MERRIMAN
Bingham Oceanographic Laboratory

C.-G. ROSSBY
University of Chicago

THOMAS G. THOMPSON
Oceanographic Laboratories University of Washington

MANAGING EDITOR

YNGVE H. OLSEN
Bingham Oceanographic Laboratory
Yale University
New Haven, Conn.

The Journal is normally issued in three numbers per year constituting one volume, except for volume V, published in 1942–1944, and volume VI, published in 1946 and 1947. All the income derived from subscriptions will be used solely to increase the volume and improve the quality of the Journal for the benefit of its subscribers and contributors.

Subscription price three dollars per volume, or one dollar per number. Address all communications to the Managing Editor. See inside rear cover for Note to Authors.

SEARS FOUNDATION FOR MARINE RESEARCH
BINGHAM OCEANOGRAPHIC LABORATORY, YALE UNIVERSITY

JOURNAL OF
MARINE RESEARCH

VOLUME VII 1948 NUMBER 2

THE MEASUREMENT OF THE VISCOSITY COEFFICIENT OF SEA WATER

By

YASUO MIYAKE and MASAMI KOIZUMI

Central Meteorological Observatory, Japan

In an old study by Ruppin and Krümmel (Krümmel, 1907: 281), the viscosity of sea water was measured by the method of Wil. Ostwald. However, at that time the data for the viscosity of sea water were given relative to that of pure water at 0° C. Since then several precise measurements (Dorsey, 1940: 182) have been made and reliable data obtained, and therefore it is necessary to determine more exact values for the viscosity of sea water.

Method of Measurements. The Ostwald viscometer was used. The size of the capillary was so designed that about 5 cc. of sea water could run through it in five to ten minutes. The temperature variation of a thermostat in which the viscometer was submerged was adjusted to within 0.01° C.

The measurements were made at temperatures near 0°, 5°, 10°, 15°, 20°, 25° and 30° C. Sea water with a chlorinity of 19.39 °/₀₀ was diluted to concentrations which were respectively 2, 4, 6, 8, 10, 12, 14, 16 and 18 °/₀₀; these waters were used as samples.

The coefficients of viscosity can be calculated by the formula

$$\eta = \eta_w \frac{\rho \cdot t}{\rho_w \cdot t_w},$$

where η is the coefficient of viscosity of sea water at any given temperature, η_w the coefficient of pure water at the same temperature, ρ and ρ_w respectively the density of sea and pure water at the given

temperature, t and t_w time (sec.) of flow of sea and pure water through the capillary of a viscometer. The time of flow was measured five times for each sample with a stop watch. Since each observed value for the time of flow agreed within 0.2 sec., the possible errors attendant on the determination of t were smaller than 0.1%. The error in the temperature due to the variation of the thermostat was also of the same order. The most important source of error might be due to a delicate change in the condition of the inner wall of the capillary, but it was impossible to estimate its exact magnitude.

The Results of Measurements. In Table I, the results of measurements are shown. Each value in this table was corrected for temperature and chlorinity. We can see that the viscosity of sea water decreases with increasing temperature and increases with chlorinity. O. Krümmel expressed the relation between the viscosity and the temperature of sea water in the following empirical formula:

$$\eta = \frac{\eta_0}{1 + \alpha\theta + \beta\theta^2},$$

Where η_0 is the viscosity of sea water at 0° C., α and β are constants varying with the chlorinity, and θ is the water temperature. We examined the above formula, using our experimental data. Since we found that the formula was sufficient for our purpose, α and β were determined with the above experimental data, where the values of η_0 were those obtained both by experiment and by interpolation.

In Table II, values of α and β for each chlorinity are shown, where values for the odd chlorinity were calculated by interpolation. Using these values of α and β, the viscosity coefficients of sea water for each per mille and each 1° C. were calculated by the above formula. Results are shown in Table III.

Since the mean deviation between calculated and experimental values is smaller than 0.25%, this table will be useful for obtaining viscosity coefficients of sea water of different chlorinity and temperature.

In conclusion, the authors wish to express their thanks to Mr. H. Honda for his earnest assistance throughout this work.

REFERENCES

Dorsey, N. E.

1940. Properties of ordinary water-substance. Amer. Chem. Soc., Monogr. Ser. No. 81. 673 pp.

Krümmel, Otto

1907. Handbuch der Ozeanographie. Die räumlichen, chemischen und physikalischen Verhältnisse des Meeres. Bd. 1. J. Engelhorn, Stuttgart. 526 pp.

TABLE I—Observed Values of Viscosity Coefficient of Sea Water
(Unit; Milli Poise)

Cl	*Water Temperature* (°C)						
(‰)	0	5	10	15	20	25	30
0	17.94	15.19	13.10	11.45	10.09	8.95	8.00
2	18.03	15.26	13.17	11.51	10.15	9.04	8.06
4	18.11	15.32	13.23	11.60	10.22	9.12	8.14
6	18.21	15.41	13.29	11.71	10.29	9.20	8.18
8	18.29	15.52	13.39	11.76	10.39	9.25	8.26
10	18.36	15.61	13.52	11.83	10.44	9.32	8.31
12	18.45	15.71	13.61	11.97	10.50	9.38	8.37
14	18.54	15.80	13.74	12.03	10.58	9.49	8.43
16	18.62	15.91	13.82	12.08	10.67	9.57	8.50
18	18.70	15.97	13.89	12.21	10.71	9.63	8.56
20	18.90	16.12	14.05	12.31	10.87	9.69	8.61

TABLE II—Values of α and β

$$\left(\eta = \frac{\eta_0}{1 + \alpha\theta + \beta\theta^2}\right)$$

Cl (‰)	α	β
1	0.03481	0.0002110
2	0.03489	0.0002024
3	0.03490	0.0001945
4	0.03490	0.0001891
5	0.03494	0.0001840
6	0.03496	0.0001825
7	0.03478	0.0001855
8	0.03447	0.0001897
9	0.03422	0.0001915
10	0.03397	0.0001923
11	0.03363	0.0002005
12	0.03325	0.0002219
13	0.03297	0.0002290
14	0.03279	0.0002314
15	0.03274	0.0002270
16	0.03272	0.0002219
17	0.03253	0.0002265
18	0.03225	0.0002360
19	0.03222	0.0002370
20	0.03239	0.0002357

TABLE III—Viscosity Coefficients of Sea Water, Calculated by the Formula, $\eta = \frac{\eta_0}{1 + \alpha\theta + \beta\theta^2}$, Using the Values of α and β in Table II. Cl, Chlorinity in Pro Mille; θ, Water Temperature; Unit, Milli Poise

	Cl (°/₀₀)																					
	0	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	
0	17.94	17.98	18.03	18.07	18.11	18.16	18.21	18.25	18.29	18.33	18.36	18.40	18.45	18.49	18.54	18.58	18.62	18.66	18.70	18.80	18.90	0
1	17.34	17.37	17.42	17.46	17.50	17.54	17.59	17.63	17.68	17.72	17.75	17.80	17.85	17.90	17.95	17.99	18.03	18.07	18.11	18.21	18.30	1
2	16.76	16.80	16.84	16.88	16.92	16.96	17.01	17.05	17.10	17.14	17.18	17.23	17.29	17.33	17.38	17.42	17.46	17.51	17.55	17.65	17.74	2
3	16.22	16.25	16.30	16.33	16.37	16.41	16.46	16.50	16.55	16.60	16.64	16.69	16.75	16.80	16.85	16.89	16.93	16.97	17.02	17.11	17.19	3
4	15.71	15.74	15.78	15.81	15.85	15.89	15.94	15.98	16.03	16.08	16.12	16.17	16.23	16.28	16.34	16.38	16.41	16.46	16.51	16.60	16.68	4
5	15.22	15.25	15.29	15.32	15.36	15.40	15.44	15.49	15.54	15.59	15.63	15.68	15.75	15.80	15.85	15.89	15.93	15.97	16.02	16.11	16.18	5
6	14.75	14.78	14.82	14.85	14.89	14.93	14.97	15.02	15.07	15.12	15.17	15.22	15.28	15.33	15.39	15.42	15.46	15.51	15.56	15.64	15.71	6
7	14.31	14.34	14.38	14.41	14.45	14.49	14.53	14.57	14.63	14.68	14.72	14.78	14.84	14.89	14.94	14.98	15.02	15.06	15.11	15.20	15.26	7
8	13.81	13.92	13.96	13.99	14.03	14.06	14.10	14.15	14.20	14.25	14.30	14.35	14.41	14.46	14.52	14.56	14.59	14.64	14.69	14.77	14.83	8
9	13.49	13.52	13.55	13.59	13.62	13.66	13.70	13.74	13.80	13.85	13.90	13.95	14.01	14.06	14.11	14.15	14.19	14.23	14.28	14.36	14.42	9
10	13.10	13.13	13.17	13.21	13.24	13.28	13.31	13.36	13.41	13.47	13.51	13.57	13.62	13.67	13.72	13.76	13.80	13.84	13.89	13.97	14.03	10
11	12.74	12.77	12.80	12.84	12.87	12.91	12.95	12.99	13.05	13.10	13.14	13.20	13.25	13.30	13.35	13.39	13.43	13.47	13.52	13.59	13.65	11
12	12.39	12.42	12.45	12.49	12.52	12.56	12.60	12.64	12.69	12.75	12.79	12.85	12.89	12.94	12.99	13.03	13.07	13.11	13.16	13.23	13.29	12
13	12.06	12.08	12.12	12.16	12.19	12.23	12.26	12.30	12.36	12.41	12.46	12.51	12.55	12.60	12.65	12.69	12.73	12.77	12.82	12.89	12.94	13
14	11.73	11.76	11.80	11.84	11.87	11.91	11.94	11.98	12.04	12.09	12.13	12.19	12.23	12.27	12.32	12.36	12.40	12.44	12.49	12.55	12.60	14
15	11.43	11.46	11.49	11.53	11.56	11.60	11.63	11.67	11.73	11.78	11.82	11.87	11.91	11.96	12.01	12.05	12.09	12.13	12.17	12.34	12.28	15
16	11.13	11.22	11.20	11.24	11.27	11.31	11.34	11.38	11.43	11.48	11.53	11.58	11.61	11.66	11.71	11.75	11.79	11.82	11.86	11.93	11.97	16
17	10.89	10.88	10.92	10.96	10.99	11.03	11.06	11.10	11.15	11.20	11.24	11.29	11.32	11.37	11.41	11.45	11.49	11.53	11.57	11.63	11.68	17
18	10.58	10.61	10.65	10.69	10.72	10.76	10.79	10.82	10.88	10.92	10.97	11.02	11.05	11.09	11.13	11.17	11.21	11.25	11.29	11.35	11.39	18
19	10.32	10.35	10.39	10.43	10.46	10.50	10.53	10.56	10.61	10.66	10.71	10.75	10.78	10.82	10.86	10.90	10.94	10.98	11.01	11.07	11.11	19
20	10.07	10.10	10.14	10.18	10.21	10.25	10.28	10.31	10.36	10.41	10.45	10.50	10.52	10.56	10.60	10.64	10.68	10.72	10.75	10.81	10.85	20
21	9.83	9.86	9.90	9.94	9.97	10.06	10.04	10.07	10.12	10.17	10.21	10.25	10.27	10.31	10.35	10.39	10.43	10.47	10.50	10.56	10.59	21
22	9.60	9.63	9.67	9.71	9.74	9.83	9.80	9.84	9.89	9.93	9.98	10.02	10.03	10.07	10.11	10.15	10.19	10.22	10.25	10.31	10.35	22
23	9.37	9.40	9.44	9.48	9.52	9.60	9.58	9.62	9.66	9.71	9.75	9.79	9.80	9.84	9.88	9.92	9.96	9.98	10.02	10.07	10.11	23
24	9.16	9.19	9.23	9.27	9.30	9.39	9.37	9.40	9.45	9.49	9.53	9.57	9.58	9.61	9.66	9.70	9.73	9.76	9.79	9.84	9.88	24
25	8.95	8.98	9.02	9.06	9.10	9.18	9.16	9.19	9.24	9.28	9.32	9.36	9.37	9.40	9.44	9.48	9.52	9.55	9.57	9.62	9.66	25
26	8.75	8.78	8.82	8.86	8.90	8.98	8.96	8.99	9.04	9.08	9.12	9.15	9.16	9.19	9.23	9.27	9.31	9.34	9.36	9.41	9.44	26
27	8.56	8.59	8.63	8.67	8.71	8.78	8.77	8.80	8.84	8.88	8.92	8.95	8.96	8.99	9.03	9.07	9.10	9.13	9.15	9.20	9.24	27
28	8.37	8.40	8.44	8.49	8.52	8.60	8.58	8.61	8.65	8.69	8.74	8.76	8.77	8.79	8.83	8.87	8.91	8.94	8.96	9.00	9.04	28
29	8.19	8.22	8.26	8.31	8.34	8.42	8.40	8.43	8.47	8.51	8.55	8.57	8.58	8.61	8.64	8.68	8.72	8.75	8.76	8.81	8.84	29
30	8.02	8.05	8.09	8.13	8.17	8.20	8.23	8.26	8.30	8.34	8.38	8.39	8.40	8.42	8.46	8.56	8.54	8.56	8.58	8.62	8.66	30

A STUDY ON THE FOAMING OF SEA WATER. PART 1

By

YASUO MIYAKE and TOMOSABURO ABE

Centeral Meteorological Observatory, Japan

In spite of the fact that many marine observers are aware that sea water has a remarkable foaming capacity, there has not been any scientific study of this subject. However, the foaming capacity may be of considerable importance in oceanography. For instance, there must be an intimate relation between the Beaufort scale and the foaming of sea water, since there is no doubt that aspects of white waves generated on the sea by winds largely depend on this effect.

It is known that a solution of inorganic electrolyte has a foaming ability, though weak and unstable. Therefore, the foaming nature of sea water may be due for the most part to inorganic material, and it is possible that organic substances such as humus also are important.

The Degree of Foaming of Sea Water. To obtain some degree of information regarding the foaming nature of sea water quantitatively, about 100 cc. of sea water were sealed in a glass ampule, foams were made by agitation with the hand, and the height of the foamy layer was then measured quickly. Since sea water foam is not stable, the height must be read as soon as the vibration has stopped. Pure sea water collected southeast of the Bonin Islands was used as a sample, after being filtered once.

The degree of foaming of sea water was compared with other organic liquid solutions such as ethyl alcohol. Results of measurements are given in Table I.

TABLE I—Comparison of Foaming of Sea Water with Other Liquid Solutions

Solution	*Concentration*	*Height of foamy layer*
Butyl Alcohol	0.13 mol.	5.2 cm.
Amyl Alcohol	0.01 mol.	3.0 cm.
Ethyl Alcohol	0.1 mol.	2.2 cm.
Acetic Acid	0.8 mol.	1.8 cm.
Sea Water	Cl; 19.4 ‰	1.7 cm.
Artificial Sea Water	Cl; 19.0 ‰	1.2 cm.
Distilled Water	.....	0

As shown in Table I, the foaming of sea water is almost the same as that of an acetic acid solution of 0.8 mol., but it is weaker than that of 0.1 mol. ethyl alcohol solution. Distilled water shows no such foaming, which is well known. Since artificial sea water also produces foaming of the same order, a main cause of foaming may be attributed to dissolved inorganic salts. However, the foaming nature of artificial sea water is always weaker than that of natural sea water, so there must be some other causes for foaming which result from organic matters such as humus or dissolved gases.

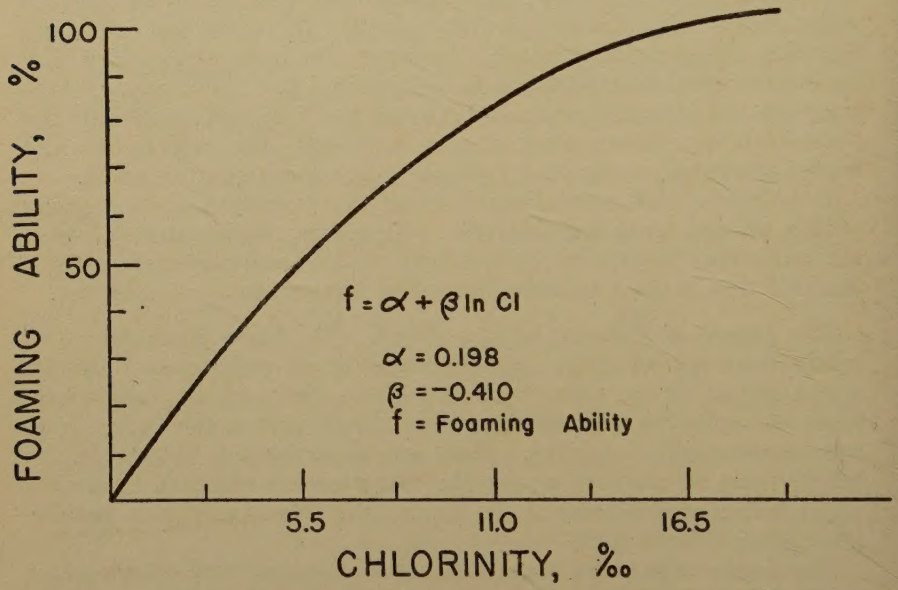

Figure 1. The relation between foaming and chlorinity.

It is also to be noted that foams of sea water are larger and more unstable than those of the solutions of organic compounds used for the comparison.

The Relation between Foaming and the Chlorinity of Sea Water. It is known that there is a relation between the foaming ability of electrolytic solutions and their concentration. In general, the degree of foaming increases with increasing concentration, and in some cases, it reaches a maximum at a definite concentration. We studied this point for sea water and obtained the results shown in Fig. 1.

In Fig. 1, relative height of the foaming layer (*i. e.*, relative foaming

ability) is given. As shown, the height of the foamy layer decreases exponentially with decreasing chlorinity. The empirical relation between the foaming layer and the chlorinity is as follows:

$$f = \alpha + \beta \ln \mathrm{Cl},$$

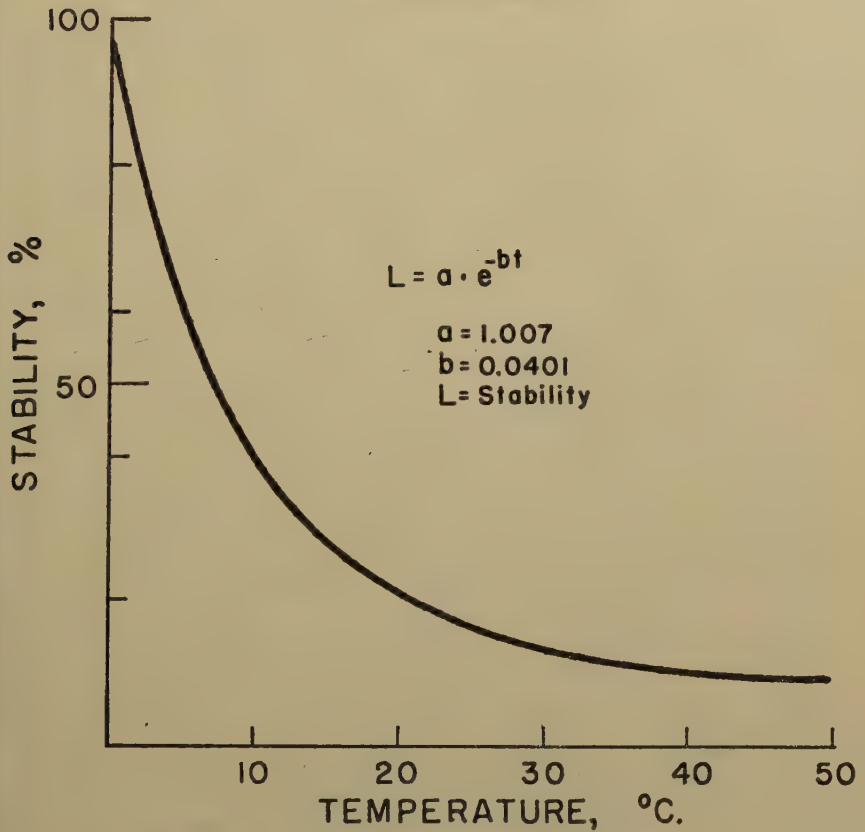

Figure 2. The relation between stability and temperature.

where f is foaming capacity (relative height of a foamy layer, sea water having Cl 19.4 ‰ as a standard), and Cl the chlorinity of sea water; α and β are constants. This formula shows that the foaming ability is proportional to the natural logarithm of the concentration of sea water.

Since the chlorinity of water with a half value of foaming ability is about 5 ‰, it can be said that the foaming capacity of sea water offshore is practically the same everywhere.

The Relation between the Stability or Life of Foams and the Temperature of Water. The temperature effect on foamy layers was studied next, and it was found that the temperature had a great influence on

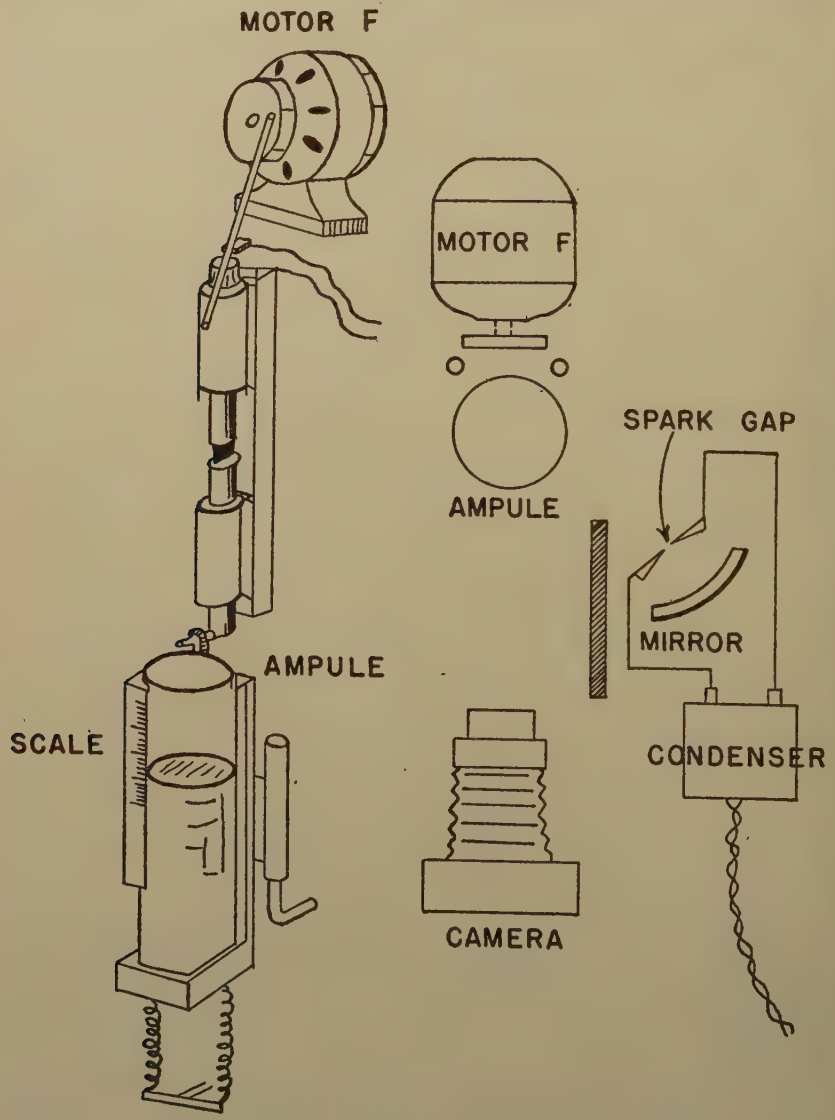

Figure 3. Diagram of apparatus.

the life of the foamy layer, but had little effect on the amount formed.

As shown in Fig. 2, the stability of a foamy layer (*i. e.*, the time of duration of the foamy layer after the vibration has stopped) has an intimate relation with the temperature of water.

The temperature was changed from 0° to 50° C. At 0° C the life was about 70 sec., it became shorter as the temperature was raised, and at 10° C the life was reduced to about half, as shown in Fig. 2.

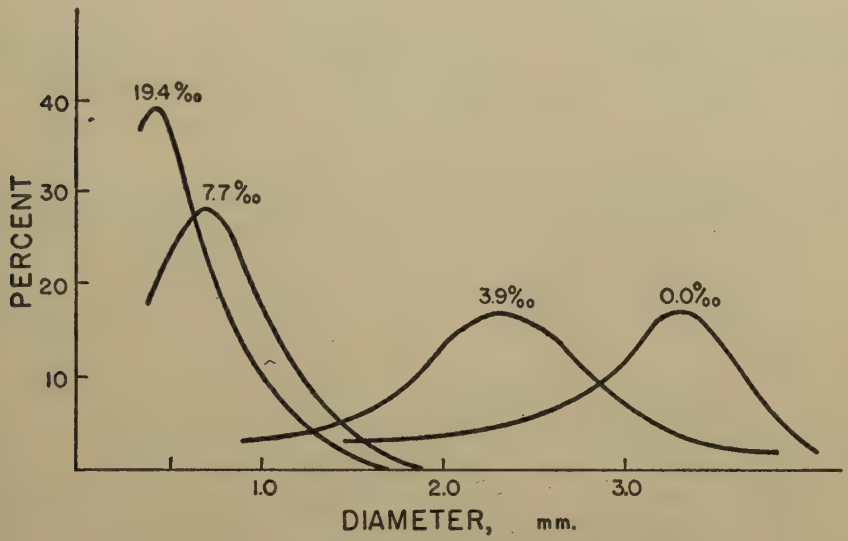

Figure 4. Distribution of the size of foam and chlorinity.

The empirical relation of stability and chlorinity is as follows:

$$L = a \cdot e^{-bt} \qquad a = 1.007, \quad b = 0.0401,$$

where L is the relative life (sea water having the chlorinity 19.4 ‰ as a standard), t the temperature, and a and b constants.

The Size of Foams. The measurement of the size of each foam bubble was obtained by means of the apparatus illustrated in Fig. 3. A photograph of the foamy layer was taken with illumination from a spark between two electrodes of aluminium.

A glass ampule was vibrated up and down by a motor (F) about fifteen times per second. As soon as the vibration ceased, a spark was flashed and a photograph was taken on a plate, which was then en-

larged and the size measured directly on the plate with a pair of dividers. Thus, the size of each bubble could be given with an accuracy of 0.2 mm. Results of measurements are shown in Fig. 4.

As shown, the distribution of size for a definite concentration forms nearly a normal distribution curve. The peak of the curve becomes lower with decreasing chlorinity, and at the same time it becomes broader.

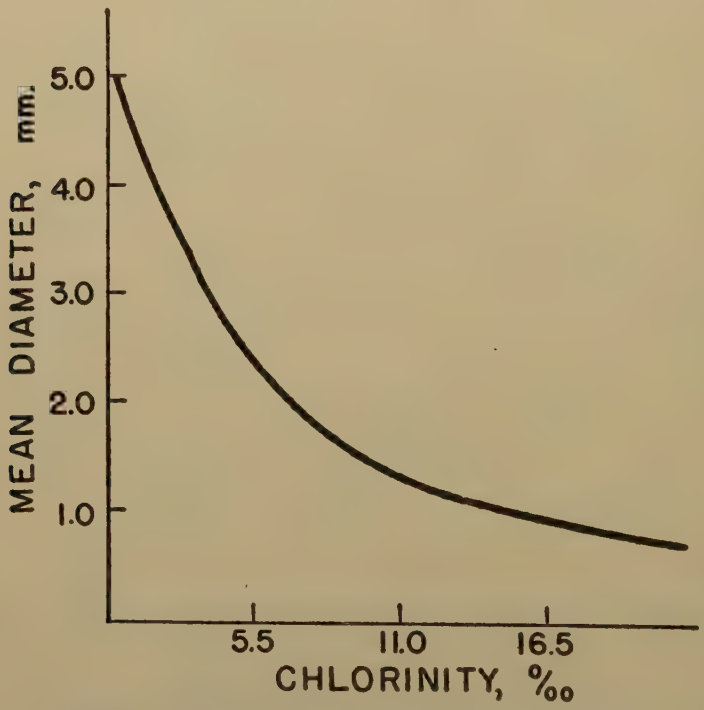

Figure 5. The relation between a mean diameter and chlorinity.

In Table II, mean diameters of bubbles for different chlorinity concentrations are given. Mean diameter $\bar{d}$ was calculated by the following formula:

$$\bar{d} = \sqrt[3]{\frac{\Sigma d^3}{\Sigma n}},$$

where d is a diameter of each bubble and n the number.

TABLE II—The Chlorinity and Mean Diameter of Foam (Bubble)

Mean diameter	*Chlorinity*
0.75 mm	19.0 °/₀₀
0.84 mm	17.4 °/₀₀
0.98 mm	15.9 °/₀₀
1.19 mm	7.7 °/₀₀
2.74 mm	3.9 °/₀₀
5.11 mm	0.0 °/₀₀

As shown in Fig. 5, the mean diameter decreases exponentially with increasing chlorinity. The empirical relation between mean diameter and chlorinity was found to be as follows:

$$\bar{d} = 5.4 \times 8.5^{-Cl}.$$

It is interesting that the foaming capacity is correlated with the inverse relation between bubble size and chlorinity. Further consideration of this effect is contemplated in the near future.

THE EQUATORIAL CURRENTS OF THE EASTERN PACIFIC AS MAINTAINED BY THE STRESS OF THE WIND[1]

By

ROBERT O. REID

Scripps Institution of Oceanography
University of California
La Jolla, California

INTRODUCTION

Sverdrup (1947) has shown that the main features of the distribution of mass in the eastern equatorial Pacific can be derived exclusively from the average wind stress distribution. In this paper the method for computing the wind stress will be developed, the treatment of the oceanographic observations will be discussed, the field of mass and of mass transport as derived from the oceanographic and wind observations will be dealt with, and finally an analytic solution will be given.

Counter currents have long been recognized from ships' records in each of the equatorial oceans, and their detailed structures in the Atlantic and Pacific Oceans have been described by Sverdrup (1932) and Defant (1936).

Montgomery (1940) and Montgomery and Palmén (1940) regarded the counter current as a downslope flow in the region of the calm belt, retarded only by frictional forces at the lateral and underlying boundaries of the current. They assumed that the hydrostatic head necessary to balance the friction was maintained by a piling up of water in the west due to the traction of the adjacent trade winds.

Stockmann (1945) has shown that it is necessary only that a sufficient lateral variation exist in the velocity of a wind of constant direction, blowing over a *closed homogeneous sea*, in order to give rise to a steady surface "compensation current," directed opposite to the prevailing wind. He has extended this theory to include the equatorial counter current developed in a *homogeneous ocean* (1946a, 1946b). In subsequent papers he includes *non homogeneous* conditions

[1] Contribution from the Scripps Institution of Oceanography, New Series, No. 353. This work represents results of research carried out for the Hydrographic Office, the Office of Naval Research, and the Bureau of Ships of the Navy Department under contract with the University of California.

(1946c), and takes lateral friction into account (1946c, 1946d), as suggested earlier by Rossby (1936a) and Montgomery (1940). In all his papers, Stockmann has neglected the fact that the Coriolis parameter varies with latitude.

Sverdrup (1947) takes into account the variation with latitude of the Coriolis parameter in his treatment of wind-driven currents in a *baroclinic* ocean, but he neglects lateral friction. His theoretical treatment, as well as that of Stockmann (1946c, d), assumes that it is possible to explain the predominant characteristics of the permanent equatorial currents and accompanying distribution of mass in a non-homogeneous sea by considering the wind stress as the only external factor explicitly involved. This would imply that the thermodynamics of the system can be ignored, and that only conservation of mass and of momentum need be considered. Such considerations evidently cannot completely account for the dynamics of the equatorial oceanic circulation, but it is intended here to show that a good approximation is achieved.

BASIC THEORY

Since we are dealing with a baroclinic medium wherein the pressure gradient generally vanishes at a moderate depth, it is apparent that horizontal motion and consequently frictional forces become negligible with depth. The net volume forces of the upper layer must therefore be balanced by the frictional stress exerted on the surface by the wind. The energy transferred to the water by the wind stress must be partly utilized by the work of cross isobaric flow and partly dissipated by turbulence in the upper layers if permanent currents are to be maintained.

Solutions in Terms of the Wind Stress. The principal assumptions, upon which Sverdrup (1947) bases his theoretical expressions, are as follows:

(a) there exists no acceleration,
(b) the fluid is free of lateral friction,
(c) the field of mass is in steady state,
(d) there exists a depth common to the whole fluid region at which horizontal and vertical motion is vanishingly small,
(e) vertical motion is absent at the surface, and
(f) the north–south component of wind stress is independent of longitude.

The above conditions lead to particular solutions of the equations of motion and continuity. The field of mass, which can be described in the horizontal by a function P, is given by the expressions

$$\frac{\partial P}{\partial x} = -R \tan\varphi \frac{\partial \tau_x}{\partial y} + \tau_x, \tag{1a}$$

and $$\frac{\partial P}{\partial y} = -(x_1 - x_0) \tan\varphi \left(\frac{\partial \bar{\tau}_x'}{\partial y} \tan\varphi + R \frac{\partial^2 \bar{\tau}_x'}{\partial y^2} \right) - 2\omega \sin\varphi \, M_{x0} + \tau_y. \tag{1b}$$

The function $P = P(xy)$ is the vertical integral of pressure defined sufficiently by

$$P \equiv \int_0^d p \, dz + \text{constant}, \tag{2}$$

where $p = p(xyz)$ is the pressure, z is the vertical coordinate measured downward from the surface, and d is the depth of no motion. In (1) x and y are the east and north coordinates, respectively, of a rectilinear system whose origin is situated on the equator; $x_0 = x_0(y)$ is the position of the eastern boundary; and $x_1 = x_1(y)$ is the locus of an arbitrary path west of $x_0(y)$. The term R is the radius of the earth, φ is the latitude (positive north of the equator), $\tau_x = \tau_x(xy)$ and $\tau_y = \tau_y(xy)$ are the east and north components of wind stress, respectively, and $\bar{\tau}_x' = \bar{\tau}_x'(y)$ is the average stress between the continent and the western path, x_1, i. e.,

$$\bar{\tau}_x' \equiv \frac{1}{(x_1 - x_0)} \int_{x_0(y)}^{x_1(y)} \tau_x dx. \tag{3}$$

In (1b) ω is the angular speed of the earth, and $M_{x0} = M_x(x_0 y)$ represents the eastward transport at the boundary.

The mass transport components are given by the expressions

$$M_x = \frac{x_1 - x_0}{2\omega \cos\varphi} \left(\frac{\partial \bar{\tau}_x'}{\partial y} \tan\varphi + \frac{\partial^2 \bar{\tau}_x'}{\partial y^2} R \right) + M_{x_0}, \tag{4a}$$

and $$M_y = -\frac{R}{2\omega \cos\varphi} \frac{\partial \tau_x}{\partial y}, \tag{4b}$$

where $M_x = M_x(xy)$ and $M_y = M_y(xy)$ are the east and north components of the net mass transport of water per unit time through a column of depth d and unit widths. In terms of the density, $\rho(xyz)$, and the velocity components, $v_x(xyz)$ and $v_y(xyz)$, the mass transport is given by the expressions

$$M_x \equiv \int_0^d \rho v_x \, dz, \quad M_y \equiv \int_0^d \rho v_y \, dz.$$

Sverdrup approximates the American west coast by a north-south

vertical boundary, in which case M_{x0} vanishes as required by the kinematic boundary condition (i. e., that the normal velocity at the boundary be zero). In general it can be shown, on the basis of (4b), that the transport for any vertical boundary takes the form

$$M_{x_0} = -\frac{R}{2\omega \cos\varphi}\left(\frac{\partial \tau_x}{\partial y}\right)_0 \tan\theta\,, \tag{5}$$

where the zero subscript refers to the boundary, and $\theta = \theta\,(y)$ is the bearing of the coast.

It can be demonstrated, however, that for the stress distribution of the *trade wind area*,

$$\left(\frac{\partial \tau_x}{\partial y}\right)_0, \quad \frac{\partial \bar{\tau}_x'}{\partial y} << \frac{\partial^2 \bar{\tau}_x'}{\partial y^2} R.$$

Furthermore, the mean value of tan θ is approximately unity. This means that M_{x0} is negligible compared with M_x if $(x_1 - x_0)$ is of the same order of magnitude as R, and since tan φ is less than unity, (1b) reduces approximately to

$$\frac{\partial P}{\partial y} = -(x_1 - x_0)\, R \frac{\partial^2 \bar{\tau}'_x}{\partial y^2} \tan\varphi + \tau_y \equiv f_2(\tau), \text{ for } |x_1 - x_0| > R\,. \tag{6}$$

Likewise, (4a) becomes approximately

$$M_x = \frac{(x_1 - x_0)}{2\omega \cos\varphi} R \frac{\partial^2 \bar{\tau}_x'}{\partial y^2} \equiv f_3(\tau), \text{ for } |x_1 - x_0| > R\,. \tag{7}$$

Near the boundary the more exact equations, (1b) and (4a), must be used.

Equations (1a) and (4b) are averaged over the range from x_1 (y) to a second arbitrary path, x_2 (y), neglecting the boundary contributions to the derivative of mean stress[2]:

$$\frac{\Delta P}{\Delta x} = -R \tan\varphi \frac{\partial \bar{\tau}_x}{\partial y} + \bar{\tau}_x \equiv f_1(\tau)\,, \tag{8}$$

and

$$\bar{M}_y = -\frac{R}{2\omega \cos\varphi} \frac{\partial \bar{\tau}_x}{\partial y} \equiv f_4(\tau)\,. \tag{9}$$

In (8) and (9) $\bar{\tau}_x = \bar{\tau}_x\,(y)$ is the mean stress between the arbitrary boundaries, defined by an expression similar to (3), where x_0 is replaced

[2] Terms of the form $\frac{\bar{\tau}_x(y) - \tau_x(x_1 y)\tau}{\Delta x} \cdot \frac{dx_1}{dy}$, thereby neglected, are of small consequence except for extreme variation of $\tau(xy)$ with x, or extreme values of dx_1/dy or dx_2/dy.

by x_2; $P = P(x_1y) - P(x_2y)$ is the difference of P for two stations at the same latitude; and $\Delta x = (x_1 - x_2)$ is the range. In (9) $\bar{M}_y = \bar{M}_y(y)$ is the mean north-south transport between x_1 and x_2.

The approximations (6), (7), (8), (9) are dependent upon the existence of but one solid boundary. Consequently, a second kinematic boundary condition to the west cannot in general be satisfied by the solutions, and application to only a limited region of the Pacific is in order. This inadequacy of the solutions is presumably due to neglect of field accelerations.

Solutions Based Mainly on Oceanographic Data. The pressure can be determined to a very close degree of approximation by the hydrostatic equation,

$$dp = \rho dD, \tag{10}$$

where $D = D(xyz)$ is the geopotential depth. Since we are ultimately concerned with differences of P, the anomaly of geopotential between the isobaric surfaces $p(xyz)$ and $p(xyd)$ can be substituted for D. Integrating (10) and making use of (2), the function P takes the form

$$P = \bar{\rho} \int_0^d (\Delta D_d - \Delta D) dz + \text{constant}, \tag{11}$$

where $\Delta D = \Delta D(xyz)$ is the geopotential anomaly between the surfaces $p(xy0)$ and $p(xyz)$, and ΔD_d is that between the surfaces $p(xy0)$ and $p(xyd)$. The term $\bar{\rho}$ is the mean density of the column of depth d.

The mass transport can be obtained from the following expressions:

$$M_x = \frac{1}{2\omega \sin \varphi} \left(\tau_y - \frac{\partial P}{\partial y} \right), \tag{12a}$$

and

$$\bar{M}_y = \frac{-1}{2\omega \sin \varphi} \left(\bar{\tau}_x - \frac{\Delta P}{\Delta x} \right), \tag{12b}$$

obtained directly from the integrated equations of motion. These equations are a good approximation for $|\varphi| > 0$. To avoid an indeterminate expression at $\varphi = 0$, equation (12a) is differentiated with respect to y and solved for M_x at $\varphi = 0$. This yields

$$M_x = \frac{R}{2\omega} \frac{\partial}{\partial y} \left[\tau_y - \frac{\partial P}{\partial y} \right], \quad \varphi = 0. \tag{13a}$$

Similarly, the north-south mass transport is

$$\bar{M}_y = -\frac{R}{2\omega} \frac{\partial}{\partial y} \left[\bar{\tau}_x - \frac{\Delta P}{\Delta x} \right], \quad \varphi = 0. \tag{13b}$$

If lateral friction and acceleration are not negligible in the vicinity of the equator, then equations (13a, b) are rigorous only if the former quantities are constant or attain an extreme value at $\varphi = 0$.

The above equations serve as an independent means of checking the expressions involving stress only.

WIND STRESS THEORY

The expression used for calculating the stress of the wind, $\tau\ (txy)$, is

$$\tau\ (txy) = \rho'\gamma^2 u^2\,, \qquad (14)$$

where ρ' is the density of the air at sea level, $u = u\ (txy)$ is the wind speed at a fixed height, and γ^2 is the resistance coefficient, a non-dimensional parameter. In general γ^2 is a function of the roughness of the surface and of the height at which the wind speed is measured. Rossby and Montgomery (1935), Rossby (1936b), and Munk (1947) have shown that a discontinuity in roughness occurs at a wind force of about 4 Beaufort. Assuming that the estimated wind velocity applies to a height of 8 meters, the values of the resistance coefficient above and below the critical wind speed are approximately:

$$\gamma_1^2 = 0.8 \times 10^{-3},\ u < \text{Beaufort } 4\,,$$
$$\gamma_2^2 = 2.6 \times 10^{-3},\ u > \text{Beaufort } 4\,.$$

These values, adopted by Sverdrup, *et al.* (1942), hold strictly for air of indifferent stability. It will be assumed here that the temperature lapse rate over the equatorial Pacific, on the average, equals that at indifferent equilibrium.

As indicated by Montgomery and Palmén (1940), because of the quadratic relationship between τ and u, the variability of the wind must be properly considered in evaluating the mean stress exerted by the wind. The complex variability of the wind velocity can be examined according to magnitude fluctuation and directional variation independently. The effect of variable magnitude is treated theoretically below.

The variability of wind speed in the *trade wind region* is assumed to be such that the distribution of *magnitude* about the mean, for any given direction and position, can be closely approximated by the Guassian law of error. At low mean speed the actual distribution probably becomes asymmetrical, which is contrary to the normal Guassian curve; but this fact is not important, since the stress exerted by a weak wind is exceedingly small.

In the following analysis, it is implied that the surface roughness is a

discontinuous function of the wind speed. The critical speed is taken as 670 cm/sec.

For a Guassian distribution, the time mean of the stress, $\tau\,(xy)$, in any specified direction, is given by the expression

$$\tau\,(xy) = \frac{\rho' \int_{-\infty}^{\infty} \gamma^2 u^2 f\, ds}{\int_{-\infty}^{\infty} f\, ds}, \tag{15}$$

where the frequency f and the parameter s are defined as follows:

$$f = \frac{1}{\sqrt{2\pi}} exp\,(-s^2/2), \tag{16a}$$

and

$$s = \frac{u - \bar{u}}{\sigma}. \tag{16b}$$

In (16), $\bar{u} = \bar{u}\,(xy)$ is the time mean of wind speed, and σ is the standard deviation of u.

The denominator of (15) is unity, so that in the special case where $\bar{u} = 670$ cm/sec, the equation takes the form

$$\tau\,(xy) = \rho' \left[\gamma_1^2 \int_{-\infty}^{0} u^2 f\, ds + \gamma_2^2 \int_{0}^{\infty} u^2 f ds \right]. \tag{17}$$

Inserting the expressions for f and u in terms of s, from (16), it follows that

$$\tau\,(xy) = \frac{\rho'}{\sqrt{2\pi}} \left\{ (\gamma_1^2 + \gamma_2^2) \left[\sigma^2 \int_{0}^{\infty} s^2 exp\,(-s^2/2)\, ds + \right.\right.$$

$$\left.\left. \bar{u}^2 \int_{0}^{\infty} exp\,(-s^2/2)\, ds \right] + (\gamma_2^2 - \gamma_1^2)\, 2\sigma\bar{u} \int_{0}^{\infty} s\, exp\,(-s^2/2)\, ds \right\}. \tag{18}$$

Upon evaluation of the definite integrals, (18) simplifies to

$$\tau\,(xy) = \rho' \left[\frac{\gamma_1^2 + \gamma_2^2}{2} (\bar{u}^2 + \sigma^2) + (\gamma_2^2 - \gamma_1^2) \frac{2\sigma\bar{u}}{\sqrt{2\pi}} \right]. \tag{19}$$

Equation (19) applies strictly to the unique distribution whose mean is force 4 Beaufort. With slight modification (19) can be applied to other distributions. If σ is small enough, say ½ Beaufort interval, then at least 97 per cent of the distribution associated with a mean wind of either Beaufort 3 or 5 lies on the lower or higher side, respectively, of the critical speed (Fig. 6). Consequently, if γ_1^2 and γ_2^2 are replaced by the appropriate single value of γ^2, (19) will yield

the stress with good approximation for any distribution whose mean is removed by at least 2σ from the critical speed.

With $\rho' = 1.16 \times 10^{-3}$ gm/cm^3 (corresponding to air at 25° C and 1000 mb pressure), the equations of stress are (for $\sigma \leq \frac{1}{2}$ Beaufort interval):

$$\tau\,(xy) = 0.9 \times 10^{-6}(\bar{u}^2 + \sigma^2), \qquad \bar{u} \leqq 3 \text{ Beaufort}, \qquad (20a)$$

$$\tau\,(xy) = 2.0 \times 10^{-6}(\bar{u}^2 + 0.85\,\bar{u}\sigma + \sigma^2), \qquad \bar{u} = 4 \text{ Beaufort}, \qquad (20b)$$

$$\tau\,(xy) = 3.0 \times 10^{-6}(\bar{u}^2 + \sigma^2), \qquad \bar{u} \geq 5 \text{ Beaufort}, \qquad (20c)$$

where $\bar{u}$ and σ are in cm/sec and τ is in dynes/cm^2.

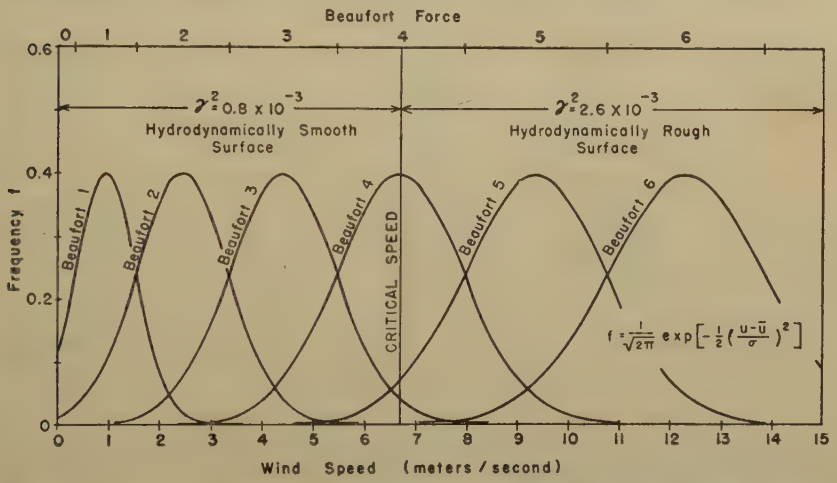

Figure 6. Assumed frequency distribution of wind speed for each integer Beaufort number of mean wind speed in the equatorial Pacific. The standard deviation of wind speed is ½ Beaufort interval. The sea surface roughness is indicated.

The value of σ was estimated for various values of wind speed in the trade wind area, and in each case it was found to be approximately ½ Beaufort interval (indicating an increase in the magnitude of σ in cm/sec with increase in mean wind speed). The wind stresses for each integer Beaufort number of mean speed in the trade wind area (Table I) have been calculated on the basis of this estimate and equations (20), using the velocity conversions of the *Smithsonian Tables*. Stresses for $\sigma = 0$ are included for comparison.

COMPUTATIONS

Because of the paucity of wind observations for the periods during which the oceanic data were obtained, it is necessary to turn to cli-

matological statistics to obtain an accurate description of the distribution of wind. The data consist essentially of sixteen and eight sector wind roses from the October and November *Pilot Charts* (U. S. Hydrographic Office, 1946) and the September-October-November *Pilot Chart* (U. S. Hydrographic Office, 1942) for the Pacific. Wind roses are given for each five degree square, such that a total of 83 are included within the area concerned.

The available oceanographic observations consist principally of two sections running roughly north–south: section A, *Carnegie* stations

TABLE I—Assignment of Wind Stress for the Trade Wind Region*

Beaufort number of mean wind	*Velocity Range (mps)*	*Velocity $\bar{u}$ (cm/sec)*	*Stress, τ, dynes/cm²*	
			$\sigma = 0$	$\sigma = \frac{1}{2}$ *Beauf. int.*
1	0.4– 1.5	95	.01	.01
2	1.6– 3.3	245	.06	.06
3	3.4– 5.4	440	.18	.19
4	5.5– 7.9	670	.89	1.06
5	8.0–10.7	935	2.65	2.70
6	10.8–13.8	1230	4.60	4.65
7	13.9–17.1	1550	7.25	7.30
8	17.2–20.7	1890	10.80	10.85

* The stress calculated on the basis of no standard deviation of wind speed has been included in the table (first stress column) for comparison with the values calculated for the trade wind area ($\sigma = \frac{1}{2}$ Beaufort interval).

147 to 159 (Fleming, *et al.*, 1945), occupied from October 15 to October 29, 1929, running from 138° 14′ W, 27° 27′ N to 159° 01′ W, 9° 24′ S; and section B, *Carnegie* stations 35 to 46, representing the period from October 26 to November 21, 1928, and *Bushnell* stations 307 to 299 (Sverdrup, *et al.*, 1943), occupied from March 18 to March 25, 1939, extending from 111° 25′ W, 22° 23′ N to 108° 20′ W, 9° 06′ S. In addition, two parallel sections are included: *Bushnell* stations 309 to 315, extending from 124° 13′ W, 29° 40′ N to 149° 35′ W, 20° 44′ N, and *Carnegie* stations 131 to 137, from 126° 20′ W, 33° 49′ N to 145° 33′ W, 24° 02′ N, for April 1939 and September 1929, respectively. Fig. 13 shows the position of these stations. Values of ΔD for each station are tabulated in the reports cited.

Wind Stress Computations. The stresses for each direction of the wind roses were evaluated by use of Table I, according to the indicated Beaufort number. Each stress was in turn weighted according to its frequency of occurrence, and broken down into east and north components. The sum of each set of weighted components divided by

the total frequency gave the components of mean stress. A summary of the computed stress components is presented in Tables II and III for October and November, respectively.

It is apparent from the change in sign of τ_y that a well defined convergence of stress is present between 7½° N and 12½° N, in agreement with the convergence of mean wind streamlines analyzed by Werenskiold (1922). Zones of maximum magnitude of stress are evident at about 5° S and 17½° N, except near the continent, where the picture is obscured by the influence of the land.

The values of mean stress $\bar{\tau}_x$ and $\bar{\tau}_x'$ were computed for the two months using sections A and B as the arbitrary boundaries x_1 and x_2,

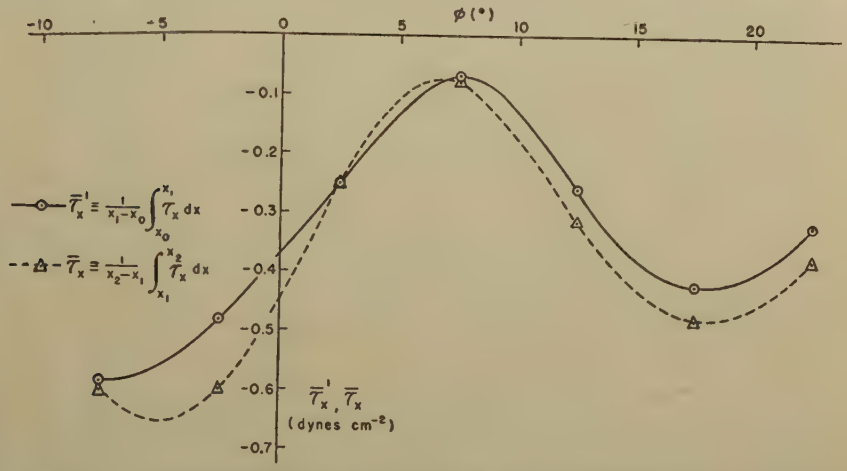

Figure 7. Latitude dependence of the mean east–west wind stress components, $\bar{\tau}_x$ and $\bar{\tau}_x'$ The curves represent climatological average stresses for the period October 1 to November 30

respectively, in the averaging process. The October-November mean wind stress distribution was taken as representative of actual conditions existing at the time of the oceanographic observations (excluding *Bushnell* data). The computed two-month averages are shown in Fig. 7. Values of τ_y were determined from the stresses along section A (for a 10° band of longitude). The October-November mean τ_y curve is given in Fig. 8.

The mean longitudinal stress curves represent a total of more than 3000 observations each (U. S. Weather Bureau, 1938). The shape of these curves suggests a sinusoidal distribution of the form

$$\tau_x = \tau_0 + a\,(\varphi - \varphi_0) + b \sin 2k\,(\varphi - \varphi_0)\,, \qquad (21)$$

which would make possible an analytic solution. This will be discussed later.

TABLE II—Summary of Computed Wind Stresses—October

τ_x and τ_y are given in the numerator and denominator, respectively, of the tabulated fractions, the units being dynes/cm².

Lat. (°)	Longitude—West														
	152.5	147.5	142.5	137.5	132.5	127.5	122.5	117.5	112.5	107.5	102.5	97.5	92.5	87.5	82.5
22.5	——	——	——	−.70 / −.39	−.66 / −.55	−.50 / −.43	−.39 / −.45	−.20 / −.30	.08 / −.13	.04 / −.03	——	——	——	——	——
17.5	——	——	——	−.79 / −.64	−.68 / −.40	−.58 / −.59	−.51 / −.34	−.30 / −.30	−.04 / −.17	.02 / −.05	−.02 / .01	——	——	——	——
12.5	——	——	——	−.63 / −.44	−.39 / −.26	−.37 / −.21	−.17 / −.11	−.06 / −.01	.08 / −.08	.10 / .00	.12 / .11	−.04 / −.32	.03 / −.01	.03 / .02	——
7.5	——	——	−.34 / .26	−.24 / .11	−.31 / .38	−.11 / .30	.08 / .27	.19 / .67	.17 / .66	.16 / .60	−.15 / .40	.35 / .23	.46 / .57	.10 / .03	.12 / .03
2.5	——	−.55 / .31	−.40 / .36	−.62 / .31	−.68 / .45	−.65 / .58	−.55 / .77	−.45 / .81	−.12 / .56	.01 / .51	−.17 / .64	.02 / .45	.14 / .58	.34 / .84	.19 / .42
−2.5	−.50 / −.09	−.50 / .04	−.73 / .14	−.49 / .21	−1.21 / .02	−.85 / .30	−.88 / .40	−.52 / .20	−.55 / .29	−.53 / .37	−.21 / .13	−.08 / .08	−.07 / .10	−.06 / .27	.05 / .17
−7.5	−.21 / .03	−.67 / −.20	−.28 / .11	−.33 / .21	−.85 / .12	−.84 / .18	−.83 / .28	−.84 / .30	−.68 / .37	−.52 / .22	−.53 / .52	−.63 / .40	−.64 / .49	−.50 / .69	−.49* / .54

* All values for −7.5° Latitude are September–October–November mean values (1942 *Pilot chart*).

TABLE III—Summary of Computed Wind Stresses—November

τ_x and τ_y are given in the numerator and denominator, respectively, of the tabulated fractions, the units being dynes/cm².

Lat. (°)	*Longitude—West* 152.5	147.5	142.5	137.5	132.5	127.5	122.5	117.5	112.5	107.5	102.5	97.5	92.5	87.5	82.5
22.5	——	——	——	−.50 / −.47	−.57 / −.33	−.44 / −.43	−.42 / −.44	−.31 / −.35	.06 / −.12	.03 / −.11	——	——	——	——	——
17.5	——	——	——	−1.03 / −.53	−.90 / −.55	−.60 / −.41	−.71 / −.53	−.58 / −.51	−.04 / −.11	.03 / −.10	.00 / −.03	——	——	——	——
12.5	——	——	——	−.69 / −.43	−.62 / −.34	−.69 / −.54	−.50 / −.21	−.54 / −.24	−.79 / −.27	−.16 / −.03	−.05 / −.03	−.29 / −.71	−.01 / −.05	−.08 / −.09	——
7.5	——	——	−.48 / .01	−.58 / .06	−.37 / .21	−.22 / .23	−.10 / .14	−.01 / .12	−.02 / .12	−.19 / .15	−.24 / .27	.01 / .25	−.08 / .15	.00 / −.05	.10 / .04
2.5	——	−.53 / .39	−.51 / .21	−.43 / .40	−.22 / .40	−.61 / .73	−.54 / .76	−.46 / .72	−.27 / .47	−.28 / .50	−.09 / .43	−.17 / .25	.07 / .38	.33 / .52	.09 / .14
−2.5	−.52 / −.06	−1.02 / −.04	−.51 / −.03	−1.05 / .14	−.79 / .34	−.89 / .33	−.81 / .47	−.76 / .41	−.64 / .31	−.29 / .06	−.08 / .09	−.06 / .08	−.02 / .06	−.04 / .16	.07 / .15
−7.5	(See Table II)														

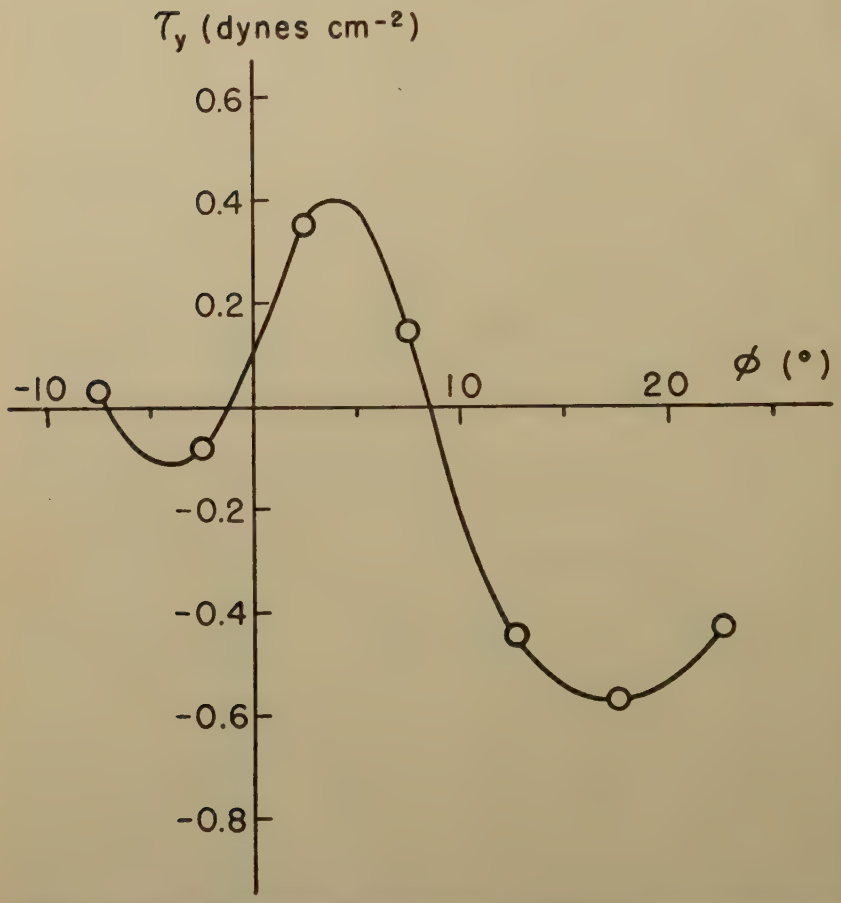

Figure 8. Plot of the north–south stress component, τ_y. The curve represents the October-November mean for a band ten degrees wide along the section A.

Using the values of $\bar{\tau}_x$ (Fig. 7), together with its numerically determined derivatives, the evaluation of the function $f_1(\tau)$ by expression (8) was carried out for each degree of latitude between 8° S and 22° N (Fig. 9). Applying (9), the function $f_4(\tau)$ was evaluated for the same range (Fig. 12). The values of $\bar{\tau}_x'$, together with the numerically determined second derivatives of $\bar{\tau}_x'$, were used to compute the stress functions $f_2(\tau)$ and $f_3(\tau)$ by equations (6) and (7), respectively (Figs. 10, 11).

Oceanographic computations. The values of P for each of the 47

stations (Table IV) were derived from equation (11) by numerical integration, using twelve standard depth intervals. The values of d and $\bar{\rho}$ were taken as 1000 m and unity, respectively, and the constant was arbitrarily set equal to zero.

A considerable discrepancy in P between the *Carnegie* and *Bushnell* stations is evident from the values of the nearly coincident runs C131 to 137 and B309 to 315 (Table IV). At a point in common with the two runs, near stations B312 and C148, it is found that the *Carnegie* value is 2.2×10^8 dynes/cm. in excess of that of the *Bushnell*.

A comparison of salinity values at great depth of the *Carnegie* cruise with those of other expeditions indicates that the former are consistently 0.03 parts per mille too low (Wüst, 1935; Schott, 1935; and Sverdrup, *et al.*, 1944). This leads to an error in P which is of the same sign and about the same magnitude as that found above.

TABLE IV—List of Stations and Integrated Pressure

Sta.	*Lat.*	*Long.*	*P*	*P′*	*Sta.*	*Lat.*	*Long.*	*P*	*P′*
	° ′	° ′	*	*		° ′	° ′	*	*
C131	33 49	126 20	57.1	0.9	B309	29 40	124 13	52.5	−1.5
132	31 38	128 48	58.6	2.4	310	28 26	128 32	54.4	0.4
133	29 21	132 30	61.6	5.4	311	27 01	132 51	57.8	3.8
134	27 45	135 22	60.9	4.7	312	25 22	137 12	59.5	5.5
135	26 39	139 07	61.3	5.1	313	23 55	141 07	60.1	6.1
136	26 13	142 02	62.1	5.9	314	22 29	144 50	60.0	6.0
137	24 02	145 33	63.2	7.0	315	20 44	149 35	60.0	6.0
147	27 27	138 14	61.1	4.9	307	22 23	111 25	54.4	0.4
148	24 57	137 44	62.0	5.8	306	20 46	109 21	52.9	−1.1
149	21 18	138 36	63.0	6.8	305	20 00	108 16	53.6	−0.4
150	16 15	137 06	56.4	0.2	304	18 34	106 23	52.8	−1.2
151	12 40	137 32	55.2	−1.0	303	17 43	105 16	54.0	0.0
152	10 05	139 44	54.3	−1.9	302AB	15 39	101 50	53.3	−0.7
153	7 45	141 24	56.7	0.5	302	13 55	99 00	58.1	4.1
154	6 42	143 22	57.8	1.6	299	9 02	86 50	52.5	−1.5
155	4 51	146 46	59.3	3.1	C 35	6 32	80 04	53.4	−2.8
156	3 01	149 46	60.1	3.9	37	5 59	82 56	55.6	−0.6
157	−1 48	152 22	58.1	1.9	38	3 46	81 37	55.6	−0.6
158	−6 33	154 58	59.5	3.3	36	2 54	80 02	58.1	1.9
159	−9 24	159 01	61.6	5.4	39	0 52	81 14	56.4	0.2
					40	−1 32	82 16	57.0	0.8
C = *Carnegie* stations					41	−1 37	86 58	54.4	−1.8
					42	−1 32	93 10	55.5	−0.7
B = *Bushnell* stations					43	−2 30	95 43	54.6	−1.6
					44	−3 15	99 48	54.2	−2.0
					45	−4 35	105 03	54.0	−2.2
					46	−9 06	108 20	56.1	−1.1

P = Integrated pressure computed directly from oceanographic data, using equation (11).
P' = Integrated pressure on an arbitrary scale, adjusted for a discrepancy of 2.2×10^8 dynes/cm between *Carnegie* and *Bushnell* data.

* All values of P and P' are given in units of 10^8 dynes/cm.

In order that the *Carnegie* data be compatible with the *Bushnell*, the quantity 2.2×10^8 dynes/cm was substracted from each of the *Carnegie* values of P. The final values of integrated pressure, P', in Table IV have been shifted to an arbitrary scale by reducing all values of P by 54.0×10^8 dynes/cm. This shift has no effect upon the gradient of P, but it makes possible a direct comparison of computed P with that found by the analytic solution to be discussed later.

A semiobjective procedure was followed in evaluating the average gradient of P in the x direction. Since the values of P for section A form a remarkably smooth curve (Fig. 10), the differences, ΔP, be-

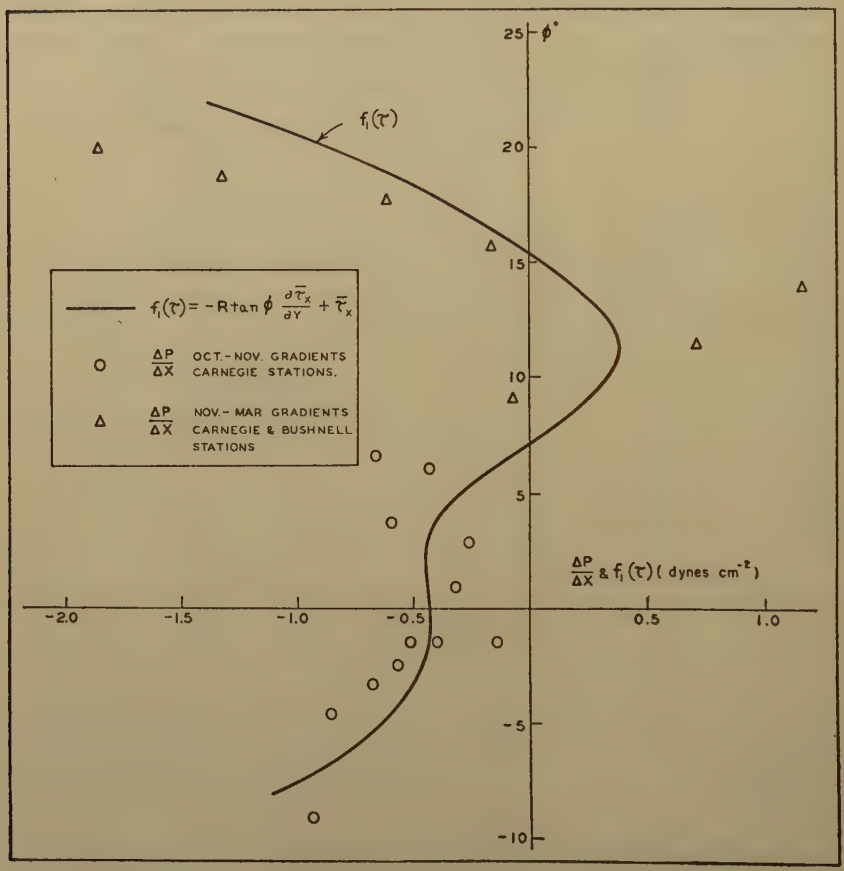

Figure 9. Plot of theoretical and observed values of $\Delta P/\Delta x$, betweeen sections A and B, as a function of latitude.

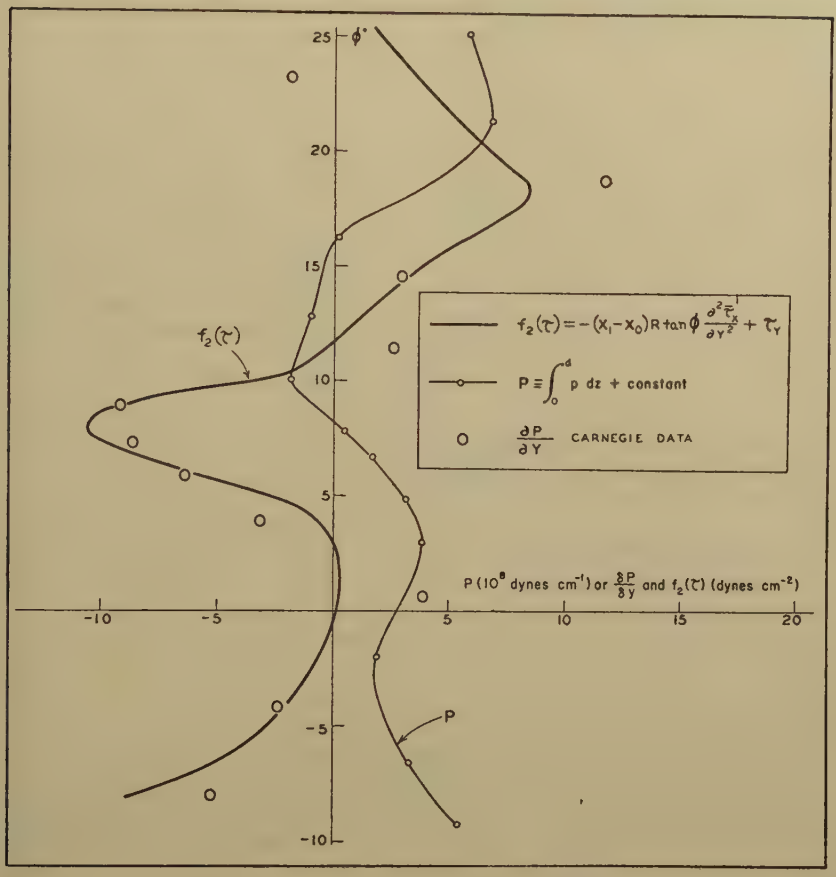

Figure 10. Plot of theoretical and observed values of $\partial P/\partial y$ as a function of latitude. Observed values have been evaluated from the *Carnegie* data along section A. Values of P for each station of section A are included.

tween the sections were computed at latitudes corresponding to those of the stations of the eastern section. Interpolation was necessary at section A only. Values of Δx were found in a similar manner, and several values of $\Delta P/\Delta x$ derived (Fig. 9).

Section A was used to obtain discrete values of the north–south gradient of P between stations. In determining the approximations of $\partial P/\partial y$, account was taken of the fact that this section is not truly north–south. The gradient is given by the following approximation:

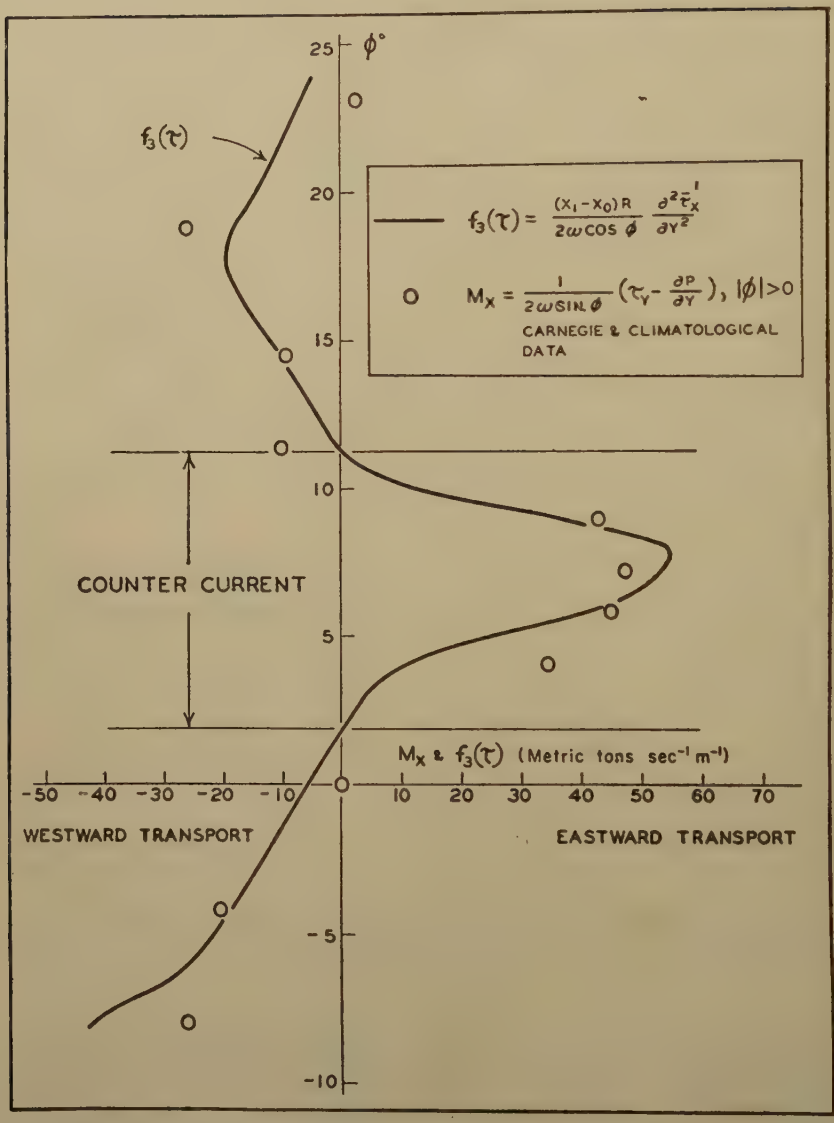

Figure 11. Latitude dependence of the longitudinal mass transport, M_x, computed by two independent methods. M_x represents the net eastward transport of water in tons per second through a column of 1000 meters depth and 1 meter width, at the *Carnegie* path (section A).

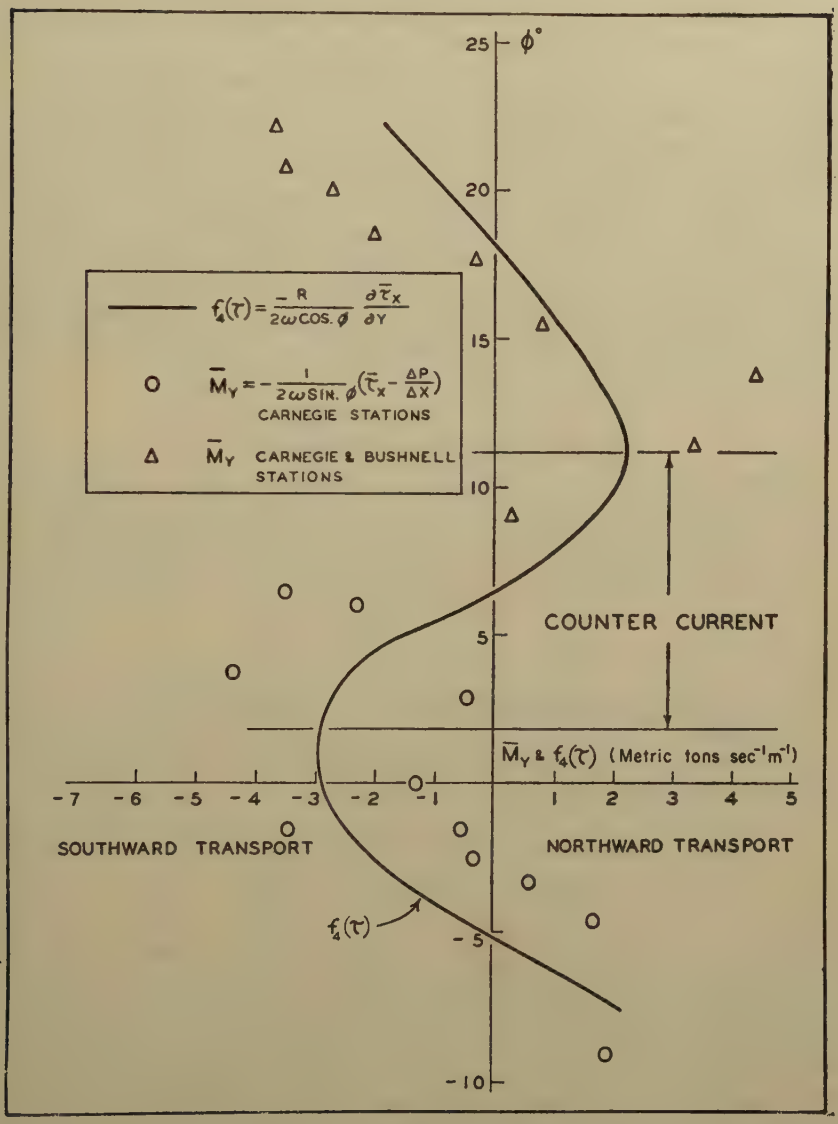

Figure 12. Latitude dependence of the latitudinal mass transport, $\bar{M}_y$, computed by two independent methods. $\bar{M}_y$ represents the mean net northward transport of water in tons per second through a column of 1000 meters depth and 1 meter width, between the two sections A and B.

$$\frac{\partial P}{\partial y} = \frac{\Delta P}{\Delta y} - \frac{\partial P}{\partial x}\frac{1}{\tan\varphi}, \tag{22}$$

where $\Delta P = \Delta P\,(x_1\, y)$ is the difference of P between adjacent stations, Δy is the corresponding change of y, and β is the bearing of the path. The gradient $\partial P/\partial x$ was approximated by $\Delta P/\Delta x$ in the determination of the final north–south gradient of P (Fig. 10).

The values of $\Delta P/\Delta x$ and $\partial P/\partial y$ for latitudes, $|\varphi|$, greater than 1° were employed in computing M_x and $\bar{M}_y$ by means of (12). Values within one degree of the equator were used in conjunction with adjacent values of the gradients to determine the second derivatives of P at the equator. This afforded means of computing the transport at the equator by (13). The resulting values are plotted in Figs. 11 and 12.

ANALYTIC SOLUTION

The essential features of the horizontal distribution of transport and of mass in the equatorial Pacific can be recognized most readily by an examination of the following simple wind stress distribution:

$$\tau_x(y) = \tau_0 + b \sin 2k\,(\varphi - \varphi_0)\,, \tag{23a}$$

and

$$\tau_y(y) = -c \sin k\,(\varphi - \varphi_0')\,, \tag{23b}$$

where τ_0, b, c, k, φ_0, and φ_0' are constants. Equation (23a) is a special case of (21).

The boundary is approximated [for the range $-15° \leqslant \varphi \leqslant 35°$] by the analytic function:

$$x_0(y) = L - my + n \sin 2K\,(\varphi - \varphi_0'')\,, \tag{24a}$$

or

$$\tan\theta = \frac{dx_0}{dy} = -m + \frac{2Kn}{R}\cos 2K\,(\varphi - \varphi_0'')\,, \tag{24b}$$

where L, m, n, K, and φ_0'' are constants.

Equations (1) and (5), together with the expressions (23) and (24), determine an analytic function $P = F\,(xy)$. The complete solution can be shown to be:

$$F(xy) = \tau_0 x + \Big[x - x_0(y)\Big]\Big\{b \sin 2k(\varphi - \varphi_0) - 2kb \cos 2k(\varphi - \varphi_0)\tan\varphi\Big\}$$
$$+ \frac{cR}{k}\cos k(\varphi - \varphi_0') + \frac{bRm}{2k}\cos 2k(\varphi - \varphi_0)$$

$$+\frac{bnK}{(k^2-K^2)}\left[K\cos 2k(\varphi-\varphi_0)\cos 2K(\varphi-\varphi_0'')\right.$$
$$\left.+K\sin 2k(\varphi-\varphi_0)\sin 2K(\varphi-\varphi_0'')\right], \tag{25}$$

provided that $k^2 \neq K^2$. The constant of integration is arbitrarily set equal to zero.

It can be further verified that the mass transport components, satisfying equations (4), (5), (23), and (24), are given by the expressions:

$$M_x = -\frac{2k^2b}{\omega R}\frac{[x-x_0(y)]\sin[2k(\varphi-\varphi_0)-\alpha]}{\cos\varphi\cos\alpha}$$
$$+\frac{kb}{\omega}\frac{\cos 2k(\varphi-\varphi_0)}{\cos\varphi}\left[m-\frac{2Kn}{R}\cos 2K(\varphi-\varphi_0'')\right], \tag{26a}$$

and

$$M_y = -\frac{kb}{\omega}\frac{\cos 2k(\varphi-\varphi_0)}{\cos\varphi}. \tag{26b}$$

In (25) and (26) the term x replaces the term x_1 in the original equations. The angle α is a small fraction of φ, according to the relation

$$\tan\alpha = \frac{\tan\varphi}{2k}.$$

For $(x-x_0) \geqslant R$, the last three terms of (25) and the last term of (26a) are unimportant.

Since the field of mass is assumed stationary, implying that the mass transport is nondivergent, there exists a stream function, $\psi = \psi(xy)$, for which

$$\frac{\partial\psi}{\partial x} = -M_y, \quad \frac{\partial\psi}{\partial y} = M_x. \tag{27}$$

In view of (26), the solution for ψ is

$$\psi = \frac{kb}{\omega}\left[x-x_0(y)\right]\frac{\cos 2k(\varphi-\varphi_0)}{\cos\varphi}. \tag{28}$$

It can be shown that (28) holds for any boundary.

The transport of water between the surface and level d is parallel to the isolines of ψ, or streamlines, and the mass flowing per unit time between adjacent streamlines is equal to the corresponding difference in ψ.

The constants were taken as follows:

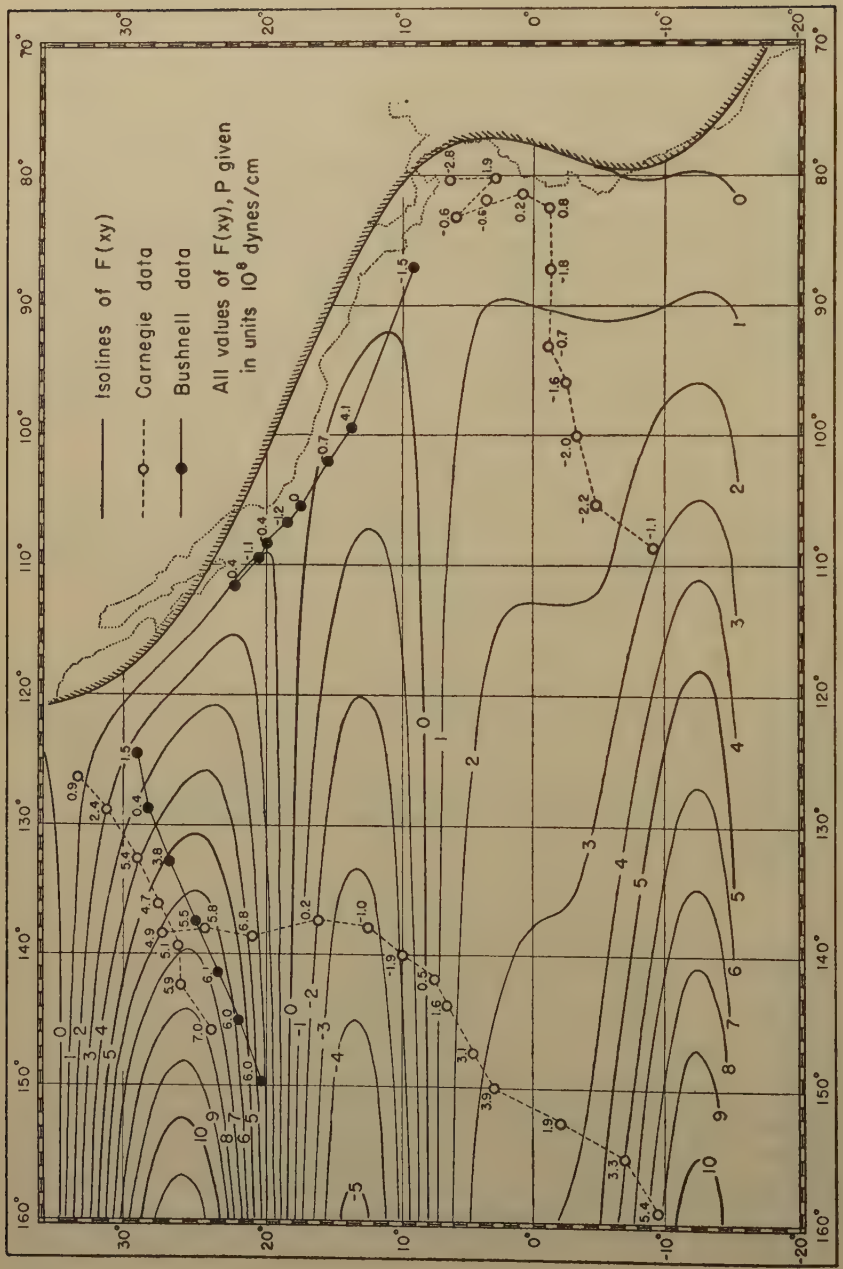

Figure 13. The horizontal distribution of integrated pressure, $F(xy)$, based on a sinusoidal distribution of longitudinal wind stress and an hypothetical boundary (see p. 92). The values of adjusted integrated pressure, P', computed from oceanographic data, are plotted for comparison.

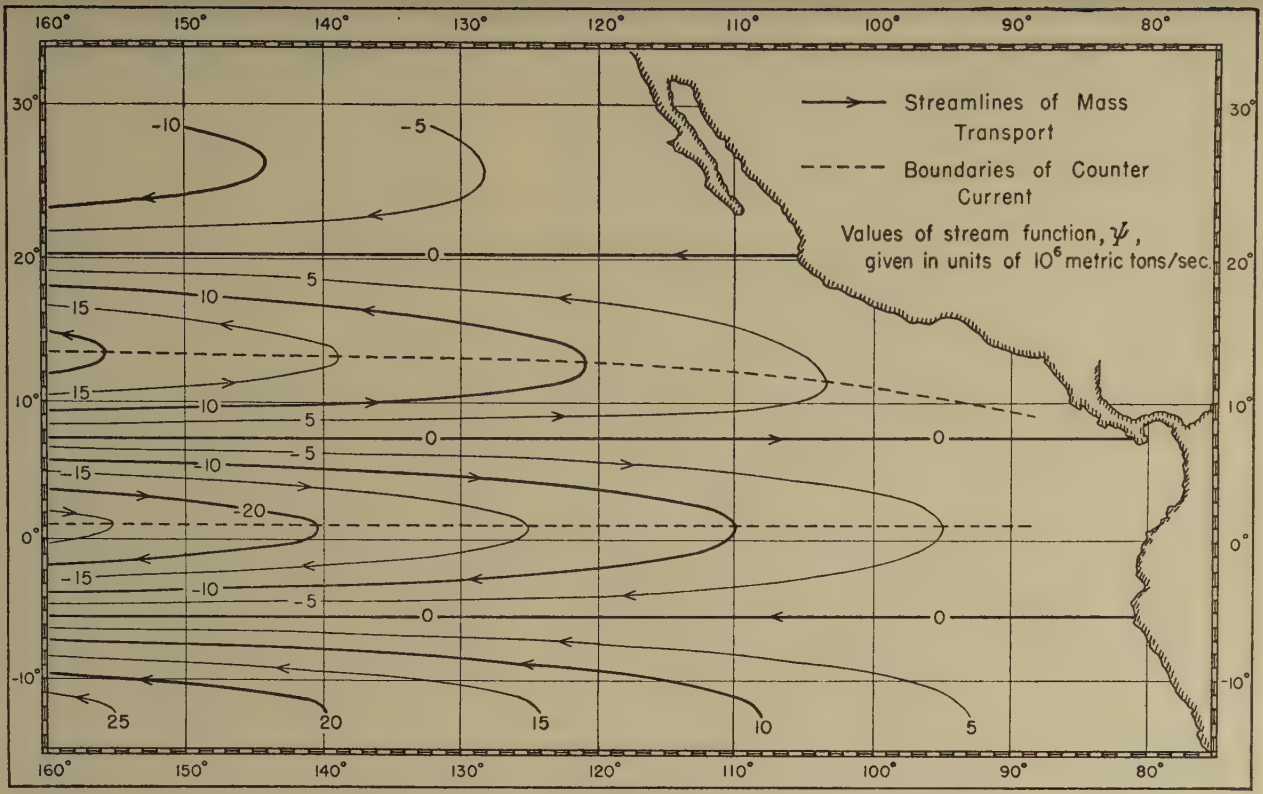

Figure 14. Streamlines of flow, representing the field of mass transport, based on a sinusoidal stress distribution and the actual boundary. Values of the stream function, ψ, are indicated. Differences of ψ between streamlines represent the net mass in millions of tons flowing per second between those streamlines (from the surface to a depth of 1000 meters). The boundary represents a streamline for which $\psi = 0$.

$\tau_0 = -0.3$ dyne/cm², $m = 0.85$, $R = 6.37 \times 10^8$ cm,
$b = 0.3$ dyne/cm², $n = 9 \times 10^7$ cm, $\varphi_0 = 1°$,
$c = 0.4$ dyne/cm², $k = 180/26 = 6.923$, $\varphi_0' = \varphi_0 + 90°/2k = 7.5°$,
$L = 80°$ W longitude, $K = 180/40 = 4.500$, $\varphi_0'' = -2°$.

Values of $F(xy)$ from (25) were determined for the region from 160° W longitude to the analytic boundary (Fig. 13). These were compared with corresponding values of P, computed from oceanic data, enabling the reduction of P to a scale in common with F (xy). The adjusted values, $P' = P - 54.0 \times 10^8$ dynes/cm (Table IV), have been plotted in the horizontal distribution chart (Fig. 13). A coefficient of correlation of 0.73 between the analytic solution, F, and the values of P' is obtained for the stations cited.

The quantity $F(xy)$ or P has a physically clear meaning in the case of the simple two layer system, where a light homogeneous layer of thickness h overlies a heavier homogeneous layer at rest. In this case, it can be demonstrated (Defant, 1936) that P is a quadratic in h. For such a system then, h would increase with distance from the coast.

Fig. 14 is a plot of the stream function based on the true boundary. It can be seen that the total mass transport of the counter current across the 140th meridian, from the difference of ψ, is about 35 million tons per second. This is somewhat higher than found for both the oceanographic and actual stress data. The boundary represents a zero valued streamline, but not an isoline of F (xy).

DISCUSSION

Figs. 9 and 10 are similar to those discussed by Sverdrup (1947). Despite the fact that the oceanic data are meager, the curves show a remarkable agreement with the principal characteristics of those observations.

The comparison of the mass transports, computed by the two independent methods, is even more striking (Figs. 11, 12). The total mass transport of the equatorial counter current, between the surface and 1000 meters computed by numerical integration of f_3 (τ), is 26.7 million metric tons per second, across the *Carnegie* section. This compares well with the estimate of transport given by Sverdrup, *et al.* (1942), which is 25 million m³ per second between the surface and 700 meters, for the same section.

The limits of the counter current are at 2° N and 12° N latitude (Fig. 11), and are indicative of the mean width of the current for the 1000 meter column. However, because of the widening of the current

with depth, indicated by the *Carnegie* profile (Sverdrup, *et al.*, 1942), the surface current should be of less width. In support of this, the October and November *Pilot Charts* (U. S. Hydrographic Office, 1946) show surface currents to the east in a band from about 5° N to 12° N.

From the plots of P (Figs. 10, 13) it is evident that the principal troughs and ridges are accounted for by the wind stress functions, except in the region of the equator, where the theory fails to describe the conspicuous rise of P to the north across the equator. It is possible that this anomolous pressure gradient could be maintained by thermodynamic processes in the vicinity of the equator. This is suggested by the fact that the observed maximum in P, just north of the equator, is almost coincident with the minimum in net evaporation given by Wüst (1936). It is clear that such a gradient, in excess of that which can be maintained by the stress of the wind at the equator, can be balanced only by lateral friction forces if unaccelerated motion is to prevail. For this reason, the picture of flow *in the vicinity of the equator* is not clear.

An alternate possibility is that the discrepancy mentioned above is not a real one. The mean atmospheric pressure distribution for October and November is asymmetrical with respect to the equator. This could bring about a discontinuity in longitudinal stress, and in turn a sharp rise of the function P across the equator. This is apparent if one considers the following approximation obtained from (1a) and (7):

$$P = (x_1 - x_0)\left(\bar{\tau}_x' - R \tan\varphi \frac{\partial \bar{\tau}_x'}{\partial y}\right) + \text{constant}. \qquad (29)$$

This expression is meaningless at $\varphi = 0$, if $\bar{\tau}_x'$ is discontinuous, but at a very small distance from the equator the expression reduces to $(x_1 - x_0)\,\bar{\tau}_x'$. Accordingly, a sharp rise of but 0.2 dyne/cm^2 in $\bar{\tau}_x'$ across the equator could lead to a corresponding rise of P in the amount of 1.0×10^5 dyne/cm along the *Carnegie* section.

SUMMARY AND CONCLUSION

By ignoring lateral friction and thermodynamic processes, it has been verified that, in the eastern equatorial Pacific, the predominant features of the distribution of mass and of mass transport of water can be accounted for semiquantitatively in terms of the wind stress alone.

The longitudinal mass transport increases nearly linearly with distance from the continent, while the north–south transport is practically independent of longitude and is a maximum very near the boundaries of the counter current.

The principal discrepancy of theory and observation exists in the proximity of the equator, where it is possible that thermodynamic processes and lateral friction are important. This discrepancy may not be real but may be brought about because of insufficient details of wind stress near the equator.

No attempt has been made to describe the vertical distribution of mass; such an analysis is to be presented in a forthcoming paper.

ACKNOWLEDGMENT

The author wishes to extend his sincere appreciation to Dr. Harald U. Sverdrup for the many suggestions and for helpful criticism which he has so generously offered during the preparation of the present paper.

REFERENCES

Defant, Albert

1936. Schichtung und Zirkulation des atlantischen Ozeans. Die Troposphäre. Wiss. Ergebn. dtsch. Atlant. Exped. *Meteor*, 6 (1), lief. *3:* 289–411.

Fleming, J. A., C. C. Ennis, H. U. Sverdrup, S. L. Seaton, and W. C. Hendrix

1945. Scientific results of cruise VII of the *Carnegie*, 1928–29. Oceanography I–B. Observations and results in physical oceanography. Graphical and tabular summaries. Publ. Carnegie Inst., *545:* 1–315.

Montgomery, R. B.

1940. The present evidence on the importance of lateral mixing processes in the ocean. Bull. Amer. met. Soc., *21:* 87–94.

Montgomery, R. B. and Eric Palmén

1940. Contribution to the question of the equatorial counter current. J. Mar. Res., *3:* 112–133.

Munk, W. H.

1947. A critical wind speed for air-sea boundary processes. J. Mar. Res., *6* (3)*:* 203–218.

Rossby, C. -G.

1936a. Dynamics of steady ocean currents in the light of experimental fluid mechanics. Pap. phys. Oceanogr. Meteorol., *5* (1)*:* 1–43.

1936b. On the momentum transfer at the sea surface. I. On the frictional force between air and water and on the occurrence of a laminar boundary layer next to the surface of the sea. Pap. phys. Oceanogr. Meteorol., *4* (3)*:* 1–20.

Rossby, C. -G. and R. B. Montgomery

1935. The layer of frictional influences in wind and ocean currents. Pap. phys. Oceanog. Meteorol., *3* (3)*:* 1–101.

Schott, Gerhard

1935. Geographie des Indischen und Stillen Ozeans. C. Boysen, Hamburg. 413 pp. (p. 203).

Stockmann, W. B.
1945. Horizontal circulation of water in the sea, as caused by non-uniform velocity distribution in a shoreward wind across its movement. C. R. (Doklady) Acad. Sci. U. R. S. S., N. S. *49* (2): 102–106.
1946a. A theory of equatorial counter-currents in the ocean. C. R. (Doklady) Acad. Sci. U. R. S. S., N. S. *52* (4): 309–312.
1946b. A theoretical explanation of certain peculiarities of the meridian profile of the surface of the Pacific Ocean. C. R. (Doklady) Acad. Sci. U. R. S. S., N. S. *53* (4): 323–325.
1946c. A tentative indirect determination of the velocities of the trade winds in the equatorial part of the Pacific. C. R. (Doklady) Acad. Sci. U. R. S. S., N. S. *53* (6): 515–518.
1946d. Equations for a field of total flow induced by the wind in a non-homogeneous sea. C. R. (Doklady) Acad. Sci. U. R. S. S., N. S. *54* (5): 403–406.
Sverdrup, H. U.
1932. Arbeider i luft- og havforskning. Beretn. Michelsens Inst. Vidensk., *2* (5): 3–20 (12–17).
1947. Wind-driven currents in a baroclinic ocean; with application to the equatorial currents of the eastern Pacific. Proc. nat. Acad. Sci., *33*: 318–326.
Sverdrup, H. U. and J. A. Fleming
1944. Scientific results of cruise VII of the *Carnegie*, 1928–29. Oceanography 1-A. Observations and results in physical oceanography. II. Results in physical oceanography. Publ. Carnegie Inst., *545*: 83–148.
Sverdrup, H. U., M. W. Johnson and R. H. Fleming
1942. The oceans, their physics, chemistry, and general biology. Prentice-Hall, Inc., New York. 1087 pp. (pp. 489–91, 631–37, 706–12).
Sverdrup, H. U. and Staff
1943. Values at standard depths, U. S. S. "Bushnell" cruise, 1939. Oceanic Obs. Scripps Inst. 1939, *1* (2): 123–128.
Werenskiold, Werner
1922. Mean monthly air transport over the North Pacific Ocean. Geofys. Publ., *2* (9): 1–55.
Wüst, Georg
1935. Schichtung und Zirkulation des atlantischen Ozeans. Die Stratosphare. Wiss. Ergebn. dtsch. Atlant. Exped. *Meteor*, 6 (1), lief. *2*: 109–288 (p. 230, footn.).
1936. Oberflächensalzgehalt, Verdunstung und Niederschlag auf dem Weltmeere. Festschrift Norbert Krebs (Länderkundliche Forschung). J. Engelhorns Nachf., Stuttgart. 368 pp. (347–359).
U. S. Hydrographic Office
1942. Pilot Chart for the South Pacific. Quarterly mean for September, October, November.
1946. Pilot Charts for the North Pacific. Monthly means for October and November.
U. S. Weather Bureau
1938. Atlas of climatic charts of the oceans. W. B. (1247): 1–130.

QUANTITATIVE STUDIES OF SUMMER PLANKTON POPULATIONS OF THE WESTERN NORTH ATLANTIC

By

GORDON A. RILEY and SAMY GORGY

Woods Hole Oceanographic Institution[1]

and

Bingham Oceanographic Laboratory

The following is an account of plankton measurements obtained during a cruise[2] of the research vessel "Atlantis" to the Mid-Atlantic Ridge between July and September 1947. It is primarily a study of plankton in the great central Atlantic eddy known as the Sargasso Sea. For comparative purposes, a small amount of material was obtained from New England coastal waters, the so-called slope water area lying off the edge of the Continental Shelf, and the Gulf Stream. An August cruise by the "Balanus" assisted in gathering this subsidiary data. The material to be described consists of phytoplankton assays by the plant pigment method of Harvey (1934), measurements of zooplankton collected by means of Clarke-Bumpus (1940) samplers, and experimental estimates of phytoplankton photosynthesis and zooplankton respiration. The plankton work was accompanied by physical and chemical measurements that will be mentioned briefly here and will be described more fully in another publication.

Before this cruise there has been little information available on the quantitative distribution of plankton in the Sargasso Sea in summer. Phytoplankton studies, consisting of cell counts (Lohmann, 1920) and pigment analyses (Riley, 1939), were available only for the May–June period. Zooplankton collections by Jesperson (1935), Leavitt (1935, 1938), and Clarke (1940) provided a fairly adequate picture of relative regional abundance, and the last author mentioned obtained enough tows at different times of the year to conclude that seasonal variations of zooplankton in the Sargasso Sea were small or nonexistent. However, all these collections were made by means of "standard" tows with unmetered nets, and the results were not readily convertible into strictly quantitative terms. It was therefore desirable to obtain a few measurements that would be more directly comparable with the

[1] Contribution No. 441 from the Woods Hole Oceanographic Institution.

[2] Jointly sponsored by the National Geographic Society, Columbia University, and the Woods Hole Oceanographic Institution.

larger body of available information on the zooplankton of the New England coastal waters and offshore banks. In addition, a first attempt has been made to learn something of the organic content and metabolic requirements of the Sargasso Sea zooplankton.

THE PHYSICAL AND CHEMICAL ENVIRONMENT

Environmental data and plankton measurements are listed in the APPENDIX. The observations were obtained near the peak of the seasonal temperature progression, of which the annual range is about 10° C in the Sargasso Sea and 20° or more in the coastal water.

Nutrients were low everywhere in the surface waters. Some of the nitrate readings were lower than the best "nitrate-free" seawater blanks that could be obtained. Probably there is a seasonal diminution in nitrate in the Sargasso Sea, since previous analyses (mostly between January and June in the western part of the area) have yielded readily measurable values of 1 to 2 mg-at per m^3. Phosphate in the surface water was very slightly higher than most of the winter and spring determinations. However, one series of phosphates obtained in November and December 1931 (Seiwell, 1935) was of the order of 0.15 mg-at of P per m^3. Thus there is some slight evidence of seasonal changes in the nutrient concentration in Sargasso Sea surface waters, but they could hardly be more than a tenth or a twentieth of the seasonal variations in temperate waters.

Minimum concentrations of nutrients generally were found at depths of 25 m to 100 m rather than at the surface, suggesting a superior rate of phytoplankton production at subsurface levels and even implying that regeneration might exceed production in the immediate surface layer. This matter will be discussed more in detail later.

PHYTOPLANKTON

Plant pigment analyses were made by filtering one to four liters of water through fine filter paper, treating the residue with acetone, and comparing the acetone solution colorimetrically with the Harvey (1934) standard.

Fig. 15 shows the surface concentration of plant pigments at twenty stations. At five Sargasso Sea stations there were sufficient analyses at other depths to estimate the total phytoplankton crop in terms of Harvey units per m^2 of sea surface, and these figures are shown in parentheses immediately below the surface values.

Analyses of coastal water ranging from 4,000 to 18,000 Harvey units per m^3 are in accord with previous studies of coastal and offshore bank

waters during the summer season. The slope water value of 11,000 units is higher than previous summer records. It suggests that the sporadic summer flowerings often noted in the coastal zone are not entirely limited to inshore waters.

In the Sargasso Sea, the following averages were obtained for the depth distribution of plant pigments: surface, 140 units; 100 m, 220 units; 200 m, 120; and 300 m, 60. The mean total population measured 49,000 units per m² of sea surface. All averages were approximately one-sixth the mean values obtained in May 1939, the only other time the method had been used in this region. It remains to be

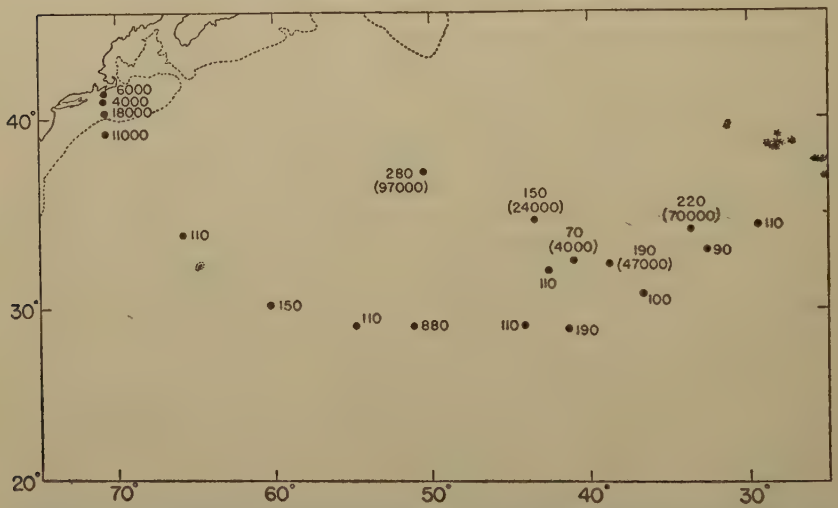

Figure 15. Phytoplankton pigment analyses. Surface concentration in Harvey units per m³; in parenthesis, total quantity in Harvey units per m² of sea surface.

seen whether this represents a normal seasonal trend or is indicative of atypical conditions at one time or the other. However, it is apparent that there was a decline during the period of the cruise, since the mean surface concentration was 180 units at stations before August 15 and 110 units after that date.

Experimental estimates of the photosynthesis of surface water phytoplankton were obtained by suspending light and dark bottles of surface water in a tub of water on deck for twenty-four hours and then measuring the differences in their oxygen content. The same method was previously used in studying a part of this area in May–June 1939 and in several cruises to Georges Bank, of which the ones in September

1939 and June 1940 are most nearly comparable from the seasonal standpoint with the present survey (cf. Riley, 1939, 1941).

Two coastal water stations of August 6 yielded values of 0.51 and 0.56 ml of oxygen produced per liter per day. These were higher than the mean (0.10) of two stations at nearly the same positions in early June 1939, but they fell within the general range of values (0.04 to 0.83 ml) obtained in the part of Georges Bank that lies between the 30 and 100 fm. curves.

A value of 0.19 ml per liter per day was found at the slope water station of August 7, as compared with a mean of 0.16 ml in two 1939 experiments and a range of 0.03 to 0.51 ml in the slope water south of Georges Bank.

Sixteen experiments in the Sargasso Sea between July 25 and September 10, 1947 ranged from −0.13 to 0.06, with a mean of −0.02. The average of previous work in May 1939 was 0.05 at five stations. It is unlikely that the mean of the recent data differs significantly from zero. However, there is little doubt that consumption exceeded production at the sea surface. The average surface water oxygen concentration was slightly undersaturated. The concentration increased between the surface and a depth of 25 m. Thus the surface content would tend to be increased both by interchange with the atmosphere and by vertical turbulence. Undersaturation could persist only under the circumstance that oxygen utilization *in situ* exceeded production.

Stations with analyses at depths of 1, 25, 50, and 100 m invariably showed an oxygen maximum at 50 m, both in absolute content and in percentage saturation (102 to 106%). At the depth of maximum concentration, which is conveniently called 50 m but might actually lie anywhere between 25 m and 100 m, vertical turbulence must continually tend to lower the oxygen concentration by mixing with undersaturated water from both greater and lesser depths. Therefore, the continued existence of a slight supersaturation at this intermediate depth must mean that oxygen production by photosynthesis exceeded the biologial utilization of oxygen.

Comparison of the Sargasso Sea stations made early in the cruise (4606, July 26; 4610, August 7) with later ones (4615, August 31, to 4617, September 4) shows no consistent trend, either seasonal or regional, in oxygen content. It therefore seems likely that a more or less steady state existed over a wide area in which the rate of increase of oxygen by phytoplankton production was approximately equal to the decrease that resulted from vertical turbulence. This can be written in the form of a simplified equation (for a more complete treatment of the problem see Sverdrup, 1938)

$$R = -\frac{A_z}{\rho} \cdot \frac{\partial^2 O}{\partial z^2}, \tag{1}$$

in which R is the net rate of production of oxygen by biological processes, A_z is the coefficient of vertical eddy diffusivity, ρ is the density of sea water, O is oxygen content, and z is depth. The vertical oxygen gradients and their change with depth can be estimated roughly from the oxygen determinations at 25, 50, and 100 m. The unknowns in the equation therefore are R and A_z. The latter can be determined by examining the change occurring during the same period in a conservative property such as temperature. The equation for the temperature change is

$$\frac{\partial \vartheta}{\partial t} = \frac{A_z}{\rho} \cdot \frac{\partial^2 \vartheta}{\partial z^2}, \tag{2}$$

in which t is time and ϑ is temperature.

TABLE I—Averages of Temperature, Oxygen, Phosphate, and Nitrate at Stations 4606, 4610, 4615, 4616, and 4617. Values Identified by (a) Are Averages of the First Two Stations, (b) the Last Three and (c) All Five

Depth in meters	25	50	100
Temperature (a)	24.8	22.0	19.5
Temperature (b)	28.0	—	20.9
Oxygen (c)	4.66	5.20	4.91
Phosphate (a)*	0.05	0.02	0.16
Phosphate (b)	0.020	0.023	0.047
Phosphate (c)	0.028	0.023	0.075
Nitrate (a)	1.95	0.40	1.15
Nitrate (b)	0.55	0.57	0.70
Nitrate (c)	1.11	0.52	0.88

* Data available only for station 4610.

Table I shows the basic data required for estimating production in the Sargasso Sea according to these equations. Phosphate and nitrate averages are included, since they will be used later for similar calculations.

The numerical approximation to equation (2) is

$$\frac{2.3}{2.77 \times 10^6} = \frac{A_z}{\rho} \cdot \frac{1}{37.5 \times 10^2} \left(\frac{-2.5}{50 \times 10^2} - \frac{-2.8}{25 \times 10^2} \right).$$

The mean temperature increase between groups (a) and (b) is 2.3°, and this divided by the difference in seconds between the median times of the two sets of observations is the time change in temperature. The temperature gradients are expressed as the difference between the 100 m and 50 m temperatures and the 50 m and 25 m ones, each divided by

the depth of the stratum in centimeters. The gradients must be expressed as the mean of stations 4606 and 4610, since 50 m temperatures were not obtained at the last three stations. The second derivative is the difference between the two gradients divided by the vertical distance between the midpoints of the two strata, namely 37.5 m. The numerical solution is $A_z/\rho = 0.19$ g per cm per second.

This figure can now be applied to equation (1) with the result that

$$R = \frac{-0.19}{37.5 \times 10^2}\left(\frac{-0.29}{50 \times 10^2} - \frac{0.54}{25 \times 10^2}\right) = 1.4 \times 10^{-3} \text{ ml of oxygen}$$

per liter per second.

The net daily oxygen production thus is estimated to have been 0.0012 ml per liter per day in the 25 m to 100 m stratum, equivalent to a carbohydrate carbon production of 0.64 mg per m^3 per day. Allowing what appears at present to be a normal phytoplankton carbon content of 17 mg per 1000 Harvey units of pigments, the estimated production represents a daily phytoplankton increase of 20% within the depth range specified. Since this depth range probably constituted nearly all of the stratum in which production exceeded consumption, it seems fair to estimate that the net total production was of the order of 48 mg of carbon per m^2 of sea surface per day, or about 6% of the total population.

No high degree of accuracy can be claimed for these results. In addition to the usual analytical and sampling errors, there is an error of unknown magnitude resulting from the exclusion of current transport and lateral mixing from the calculation. Nevertheless, in two respects the results seem reasonable. It will be shown later that the food requirements of the zooplankton appear to be slightly larger than the estimated plant production, and this is a necessary concomitant of the observation that the phytoplankton was decreasing slowly during the period of investigation. In the second place, the calculated production coefficient is of the same order of magnitude as the one previously estimated for the May 1939 data. At that time it was calculated from field experiments that photosynthetic activity produced about 4.2 g of carbon per m^2 per day, and phytoplankton respiration utilized perhaps 3.0 g. The difference of 1.2 g, or 450 mg of carbohydrate carbon per m^2 per day, which theoretically represents phytoplankton growth, is estimated to be equivalent to a growth coefficient of 8.5% of the total population per day.

Table I shows nitrate and phosphate minima at 50 m, or gradients increasing with depth. Both types of distribution are indicative of transport of these substances into the 25 m to 100 m stratum. The data are not entirely suitable for calculations of nutrient utilization according

to the principles of equation (1), since the gradients changed between the early and later part of the cruise. Therefore the existence of a steady state is questionable, and while the calculations are included for the sake of completeness, no attempt will be made to interpret their meanin .

Estimates based on the three sets of average gradients in Table I, (a), (b), and (c), show an apparent variation in both nitrate and phosphate utilization of greater than ten-fold magnitude. Thus, net phosphate consumption is found to range from 0.0005 mg of P per m^3 per day (b) to 0.0055 mg (a), and nitrate from 0.0035 mg of N per m^3 per day (b) to 0.047 mg (a). Even the largest estimates are smaller than would be required to give the phytoplankton a "normal" carbon : nitrogen : phosphorus ratio. Data by Fleming (1940) indicated an average C : N : P ratio in temperate water plankton of 41 : 7.2 : 1. The nearest approach to this that can be postulated with the present set of data is 130 : 8.5 : 1. It is fruitless to speculate on whether this implies actual nutrient deficiency in the Sargasso Sea phytoplankton or simply results from vagaries of the method.

ZOOPLANKTON

Zooplankton was collected by small, quantitative plankton samplers of the type described by Clarke and Bumpus (1940), which contain current meters to measure the volume of water strained. Oblique tows were made at approximately 2 knots for a half hour period, using nets of no. 10 silk. Usually two samplers were used at the same time, one with 0 to 400 m of wire out, the other from 400 m to 800 m of wire. In some cases only the upper sampler was used, or the lower one failed to open properly at depth. In coastal waters a single sampler was used, towing from the surface to as near the bottom as practicable.

The amount of water strained during a tow averaged about 8 m^3, with a total variation of 3 to 25 m^3. The smallest value was in coastal water where some clogging occurred. The largest was obtained from the deep sampler at a Gulf Stream station, where presumably the vertical current velocity pattern was such that the sampler towed faster than 2 kt. These are extreme cases, but they illustrate the difficulty of comparing so-called "standard" tows made with unmetered nets.

The Sargasso Sea collections generally had a volume of only a few tenths of a cubic centimeter, which was not considered adequate for volume measurements of the kind used by most previous investigators. It was desirable to use a method that would obtain greater accuracy with small samples and at the same time to maintain continuity with previous work. Therefore the samples were first prepared as for

measurements of displacement volume by draining thoroughly on a bolting silk filter after two washings with distilled water. The samples were then weighed. In a few cases where the quantity of material was sufficient for comparison of volume determinations with weights (1 to 14 cc), the volume in cubic centimeters did not differ by more than 10% from the wet weight in grams. After this the samples were dried and weighed, ignited, and weighed again. Thus the fundamental measurements obtained were wet weight (or volume) of plankton containing a small amount of trapped water, dry weight, an approximation of the weight of organic matter, and the weight of

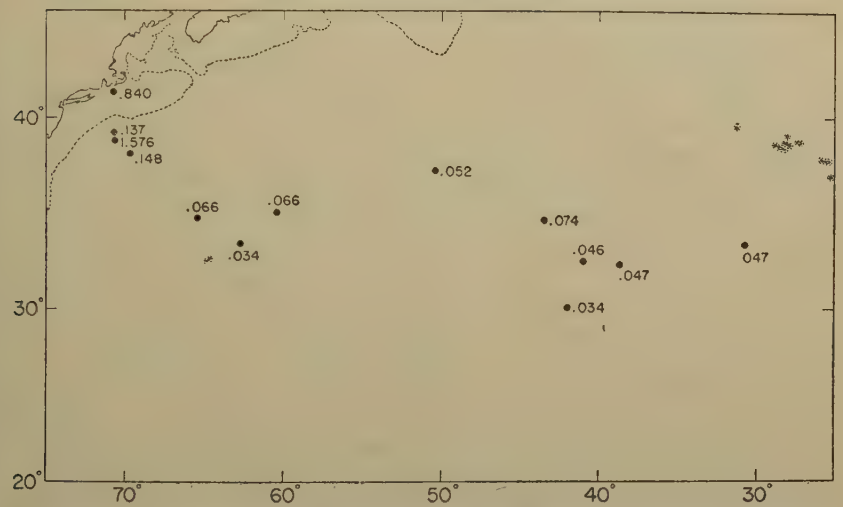

Figure 16. Zooplankton weight determinations. Wet weight in grams per m³; collections obtained by oblique tows in approximately the upper 400 m in deep water, and surface to near bottom in the coastal zone.

ash plus any sea salt that remained entrapped between the animals after the washing process. Clearly this is a compromise between the desire to obtain measurements comparable with the older work and the need for the development of more adequate quantitative methods.

Fig. 16 shows the distribution of zooplankton plotted as wet weight in grams per m³ of water in the upper stratum sampled. In most cases it is the average of slightly less than 400 m of water, the depth depending on the wire angle, but it also includes one slope water station where a single tow was made to 600 m, as well as a coastal water station where the total depth was 40 m, and the wire was let out 35 m.

The mean concentration in the nine Sargasso Sea hauls was 0.045 g per m³. The Gulf Stream station had a value about three times as high as the Sargasso Sea average. The two slope water stations had concentrations of 0.14 and 1.6 g per m³, respectively. Salps bulked large in the slope water collections. The nontunicate population was 0.025 and 0.035 g per m³, respectively. The coastal water station had a value of 0.84 g per m³, about the same as the mean of the slope water stations and nearly twenty times as much as in the Sargasso Sea.

Seven tows were made in the deeper stratum, with 400 m to 800 m

Figure 17. Zooplankton weight determinations. Rough estimate of the wet weight in grams of the total population underlying a m² of sea surface.

of wire out. Five collections in the Sargasso Sea ranged from 0.008 to 0.026 g per m³, with a mean of 0.014 g. The Gulf Stream station yielded 0.029 g per m³, and one tow in the slope water contained 0.861 g per m³. Thus in different parts of the region examined, the quantity of plankton in the deeper stratum was one-fifth to one-half the concentration nearer the surface.

The quantities that have been listed do not constitute the total zooplankton population, first because presumably some of the larger and more active animals can avoid capture by small nets, and second because Jesperson (1935) and Leavitt (1935, 1938) showed that in the slope water, in the Gulf Stream, and in the Sargasso Sea some plankton can be found to depths of 2,000 m or more. However, it was not practicable to attempt studies of the whole population, and the part

sampled is believed to be the most important from the standpoint of phytoplankton-zooplankton relations, which was the main object of the study. Therefore it seems worthwhile to examine the areal distribution of zooplankton (wet weight in g per m^2 of sea surface) obtained by the simple process of multiplying the mean concentrations by the depth towed. In this calculation the dubious correction for wire angle is not attempted in view of other errors that are unavoidable.

Fig. 17 shows the results. It is apparent that the total zooplankton was nearly the same in the coastal water and in the Sargasso Sea, while in the Gulf Stream and slope water it was considerably higher.

By way of comparison with previous work, Bigelow and Sears (1939) estimated that coastal water plankton in this area averages 0.5 to 0.8 cc per m^3. Clarke (1940) obtained an average of 0.54 cc per m^3 in 37 plankton sampler tows. He also made sampler tows in the slope water and Sargasso Sea, but the measurements were not listed in a way that permits precise comparison with present material. However, he described a series of scrim net hauls that yield an estimate of relative abundance. It was found that the mean annual ratio of plankton concentration in three areas—coastal water, slope water, and Sargasso Sea—was 16 : 4 : 1 for total zooplankton and 7 : 2.5 : 1 for the crustacean fraction.

These ratios and absolute values are more or less in accord with the more recent data. Calculations of total plankton are also of the same order of magnitude. Clarke's coastal water average of about 0.5 cc per m^3 was obtained in water of 40 m to 200 m depth, so that the total population is estimated to range from 20 cc to 100 cc per m^2 of sea surface. His estimate that the volume of the Sargasso Sea zooplankton is one-sixteenth to one-seventh the volume of the coastal crop would suggest in the former region an absolute volume of perhaps 0.03 to 0.08 cc per m^3. These are averaged over a depth range of about 900 m, indicating a total population of about 27 cc to 72 cc per m^2.

Returning to the more recent observations, rather different results were obtained on examination of the distribution of zooplankton organic matter. Fig. 18 shows that the coastal water was richer than the Sargasso Sea by a factor of three or four to one, and the Gulf Stream and the mean of the slope water stations were intermediate. These differences in distribution result from marked regional variations in the percentage of organic matter in the organisms. In the sample of coastal water plankton, the difference on ignition was 16.2% of the wet weight. By contrast, the slope water samples, consisting chiefly of salps, averaged 0.7%. The nontunicate fraction of the slope water population had an organic content of 5.8%, and in the Gulf Stream and Sargasso Sea it was respectively 5.1% and 4.8%.

The value obtained in coastal water falls within the rather large range of published values for zooplankton organic content. For example, Nakai (1942) found a range of 11% to 24% in analyses of nine species of crustacea, while Bogorov (1934) and Marshall, Nicholls, and Orr (1934) reported that the dry weight of *Calanus finmarchicus* varied from 15% to 50% of the wet weight.

The figures obtained from the slope water, Gulf Stream, and Sargasso Sea collections all were so much lower that they warranted further investigation. Material from qualitative surface tows was used for

Figure 18. Zooplankton weight determinations. Estimated organic content of the total zooplankton population in grams per m² of sea surface.

analysis. First there was the question of the degree of compaction of the small Sargasso Sea samples and the amount of extraneous water that was entrapped with the animals in the so-called wet weight measurements. Four plankton suspensions of varying concentration were prepared in the usual way. Their wet weights were determined, and the general range was from 0.05 to 1.5 g. The samples were then blotted on filter paper and weighed again. The loss ranged from 20% to 30%. The organic matter averaged 7.1% of the second wet weight measurements, which still was considerably less than the published values.

Next, a few of the major constituents were measured separately. Salps, sagittae, euphausids, and large copepods were picked out individually. Nearly pure samples of pteropods were obtained by

decanting the rest of the catch. By pouring a plankton suspension through no. 2 silk and then through no. 10, a sample was caught on the no. 10 filter that consisted almost entirely of small copepods and juvenile crustacea. Again the excess water was removed by rolling the animals on filter paper, and the samples were then treated in the usual way, with the results shown in Table II.

TABLE II—Wet and Dry Weight, Organic Matter, and Ash of Representative Plankton Groups. Tunicates Collected in Slope Water, Others in Sargasso Sea

	Weight of sample in grams				*% of wet weight*		
	Wet	*Dry*	*Org. M.*	*Ash*	*Dry*	*Org. M.*	*Ash*
Tunicates	17.3678	0.2936	0.1092	0.1844	1.7	0.6	1.1
Pteropods	1.4496	0.5590	0.0969	0.4621	38.6	6.7	31.9
Euphausids	0.1403	0.0150	0.0130	0.0020	10.7	9.3	1.4
Sagittae	0.0469	0.0063	0.0046	0.0017	13.4	9.8	3.6
Copepods	0.1058	0.0184	0.0156	0.0028	17.3	14.7	2.6
Misc. small crustacea	0.4219	0.0404	0.0310	0.0094	9.5	7.3	2.2

Clearly it is not possible to ascribe the generally low organic content of the subtropical plankton to a preponderance of soft-bodied and watery animals such as salps and medusae. Of all the groups measured, only the large copepods had what has been considered to be a normal organic content. The reasons for this situation have not been investigated further, but the data by Nakai (1942) on the chemical composition of nine species of crustacea (seven copepods, one euphausid, one amphipod) suggest a simple solution. He found that the fat content of the crustacea was highly variable, ranging from 0.4% to 12.3% of the wet weight. The nonlipoid organic matter varied from 7.8% to 14.7% of the wet weight, and it will be noted that this is approximately the range of the total organic content of the Sargasso Sea crustacea. Nakai's collections were made in various locations in the Japan Sea between 32° and 42° N and 128° to 135° E. The southern, warm water animals generally had less fat than the northern ones. Thus the fat content of five samples taken between 32° and 35° N ranged from 0.4% to 3.0% of the wet weight, with a mean of 1.3%, while eight samples from 38° to 42° N ranged from 2.1% to 12.3%, with a mean of 6.9%. There was no overlapping of species in the regional groups, and hence no assurance that the observed differences were not simply peculiarities of the species in question. Nevertheless, there is a logical inference awaiting further proof that the scarcity of phytoplankton in the southern Japan Sea and the Sargasso Sea does not favor the accumulation of fatty reserves by the zooplankton and that this is responsible for the low content of organic matter.

ZOOPLANKTON RESPIRATION

A measurement was made of the oxygen consumption of coastal water zooplankton, using a part of the plankton sampler catch mentioned above. In the Sargasso Sea, eleven similar experiments were performed with material obtained from qualitative surface tows. The general method was to stir the plankton collection vigorously and remove for experimental purposes a sample of 10 to 30 ml, the size depending on the density of the catch. The sample was then put in a 350 ml oxygen bottle, and the bottle was filled with surface water and suspended in a tub of water on deck for twenty-four hours, after which the oxygen was measured and compared with the initial content of the surface water. In most cases there was also a control bottle of surface water exposed during the same period, permitting a correction for the respiration of organisms in the water added. The zooplankton not used for experimental purposes was preserved for later study.

In one respect, the coastal water and Sargasso Sea experiments were not strictly comparable. The former was done in the dark, the latter in clear glass bottles. According to the work on *Calanus finmarchicus* by Marshall, Nicholls, and Orr (1935), daylight approximately doubled respiration. However, this is not necessarily true of Sargasso Sea zooplankton. Several light and dark bottle experiments were performed with no. 20 net catches. They were originally intended as measurements of phytoplankton photosynthesis, but the quantity of net phytoplankton was so small that essentially they were experiments on zooplankton respiration. In this series the oxygen consumption in the light bottle averaged only 16% higher than in the dark.

TABLE III—Oxygen Consumption by Sargasso Sea Zooplankton

Date	*Oxygen —ml per liter—*				*Mg C*	*—Zooplankton —Carbon consumed—*			
	0 hrs.	*24 hrs.*	*Diff.*	*Control*	*consumed per day*	*weight— (in mg) Wet*	*Org. M.*	*% of Wet wt.*	*% of Org. M.*
30 VII	4.31	2.04	2.27	−.05	0.292	102	1.98	0.29	14.8
7 VIII	4.56	3.38	1.18	−.09	0.143	84	5.11	0.17	2.8
9 VIII	4.54	3.24	1.20	−.15	0.138	30	0.93	0.46	14.8
15 VIII	4.61	4.17	0.44	−.08	0.047	13	0.43	0.37	10.9
22 VIII	4.65	1.75	2.90	—	0.354	105	3.82	0.34	9.3
24 VIII	4.64	3.91	0.73	−.31	0.055	18	0.68	0.31	8.1
26 VIII	4.61	2.96	1.65	−.20	0.190	36	1.09	0.53	17.4
1 IX	4.21	2.24	1.97	−.10	0.245	25	1.43	0.97	17.1
2 IX	4.29	2.95	1.34	−.10	0.162	20	0.97	0.81	16.7
4 IX	4.17	2.48	1.69	—	0.210	87	5.27	0.24	4.0
10 IX	4.44	1.64	2.80	−.07	0.358	41	2.24	0.86	16.0

Details of the Sargasso Sea experiments are presented in Table III. They are listed only by date, but this is sufficient to identify the posi-

tion of the collection and correlated data by reference to the material in the APPENDIX. The first two columns show the initial concentration of oxygen in the experimental bottles and the concentration at the end of the twenty-four hour period. The difference, shown in the next column, is first corrected according to the control reading and then converted to an estimate of milligrams of carbon consumed per day by the zooplankton sample, assuming a respiratory quotient of unity. Zooplankton weights are listed as the quantities estimated to have been present in the experimental bottles. The actual quantities weighed were ten to fifty times these figures. Finally, the estimated respiratory requirements (carbon consumption) are calculated as percentages of the wet weight and organic content of the zooplankton.

The results are seen to vary widely. The average figures for daily carbon consumption are 0.49% of the wet weight and 12% of the organic matter. At the coastal station the carbon consumption was found to be 0.6% of the wet weight and 4% of the organic matter. Thus the respiratory rate per unit of zooplankton organic matter appears to be lower in the coastal water than in the Sargasso Sea, but with respect to wet weight the situation is reversed. The lack of agreement in the two kinds of measurements results from differences in the organic content of the two populations.

In view of the observed variability of the Sargasso Sea experiments, not much reliance can be placed on a single coastal water observation. However, the published data on *Calanus finmarchicus* provide an estimate of metabolic requirements that is probably more or less applicable to temperate water plankton in general. Marshall, Nicholls, and Orr (1935) estimated on the basis of oxygen consumption experiments that the food requirements of Calanus in summer were of the order of 2.5% to 7.6% of their organic content. Clarke and Bonnet (1939) obtained results that averaged considerably higher, although the range of variation of individual experiments overlapped considerably. In terms of daily carbon consumption, the values established by the two groups of investigators probably lie between 1.3% and 6.7% of the organic content of the animals. The single experiment on the mixed coastal water population falls approximately in the middle of this range, but the general level of the Sargasso Sea experiments is much higher. On the other hand, the food requirement per unit of wet weight (or volume) would appear to be more or less the same in both regions.

PHYTOPLANKTON-ZOOPLANKTON RELATIONSHIP

The data that have been presented on zooplankton organic content and metabolic requirements provide a means for instructive speculation on food problems. In the Sargasso Sea, with an average content of 2.2 mg of zooplankton organic matter per m^3, the average food requirement is estimated to be 0.26 mg of carbon per m^3 per day. This probably is equivalent to about 15 Harvey units, which amounts to a daily consumption of 8% of the surface phytoplankton population, and the total zooplankton population would require 20% of the total phytoplankton crop. In order to obtain the required amount of food, the daily filtration of a quantity of zooplankton containing 1 mg of organic matter (assuming 100% removal of phytoplankton) would be 2.2 liters. In the coastal zone, an average estimate of the zooplankton carbon requirement is 5.3 mg per m^3 per day, or 6% of the surface crop of phytoplankton. In order to meet its metabolic demands, a quantity of zooplankton equivalent to 1 mg of organic matter would need to filter only 70 ml of water per day, or about one thirtieth as much as the Sargasso Sea zooplankton. Thus either the latter feeds at a much faster rate, or, if the rates are more nearly equal, the coastal water zooplankton obtains far more food than is needed to satisfy its immediate demands.

This difference in what might be called the apparent minimum filtration coefficient may not actually be as large as it appears. In every respect the conditions of the Sargasso Sea investigation would be expected to lead to maximal results. The experiments lasted only a short time, and they were done in the light and at approximately surface water temperatures. At most they are applicable to immediate surface conditions, and the mean metabolic requirement for the whole population might be only a half or a third as much (7% to 10% of the phytoplankton population per day). Such a correction probably would bring the filtration coefficient down approximately to the level of the maximum feeding rates observed by Harvey (1937) in experiments with *Calanus finmarchicus*.

Supposing that gross activity, including feeding, varies more or less in accordance with the metabolic rate (Zeuthen, 1947, has presented some evidence favoring this view), there must be a great discrepancy in the food-gathering capacity of the two populations. Judging by the experiments on oxygen consumption and the ecological considerations presented above, the feeding rates may be about the same in the two areas, and at most they average three times as high in the Sargasso Sea. If the latter population was obtaining just enough food to permit survival, it seems likely that the coastal water zooplankton was con-

suming ten to thirty times as much food as it needed to satisfy its minimum requirements. Thus its apparent superiority in fatty reserves would be accounted for readily, but the differences in food gathering capacity are so large that it would also be necessary to postulate a vastly superior rate of herbivore reproduction in the coastal zone. According to the principles of dynamic balance, this would also imply a correspondingly higher rate of predation by the larger members of the food chain.

Admittedly there is some danger in attempting to arrive at broad conclusions without examining more of the seasonal cycle. A program of investigation is now underway which will provide more critical and comprehensive data on the plankton populations of the western North Atlantic.

SUMMARY

1. The material presented includes measurements of temperature and nutrient salts, analyses of phytoplankton pigments, wet weight and organic content of quantitative zooplankton collections, and experimental determinations of phytoplankton photosynthesis and zooplankton respiration. The observations were obtained at various positions in the western North Atlantic between New England and the Azores during the period of July to September, 1947.

2. The phytoplankton populations of the coastal and slope waters yielded pigment concentrations of 4,000 to 18,000 Harvey units per m^3, which is more or less normal for this season. In the Sargasso Sea the mean surface concentration was 140 units, and the total quantity averaged 49,000 units per m^2 of sea surface. Both figures are approximately one-sixth of the only previous measurements in this region, in May 1939.

3. Experimental measurements of the photosynthesis of surface water phytoplankton yielded values of 0.19 to 0.56 ml of oxygen per liter per day in the coastal zone and slope water. Oxygen changes in Sargasso Sea surface water were not significantly different from zero.

4. The observed vertical distribution of oxygen in the Sargasso Sea was such as to indicate that biological oxygen consumption exceeded production in the immediate surface layer and that production was dominant at a depth of around 50 m. By standard physical oceanographic methods it was estimated that there existed during the period of the investigation an average net oxygen production of 0.0012 ml per liter per day in the depth range between 25 m and 100 m. On the basis of these figures, a phytoplankton growth rate of 20% per day was postulated for the depth range specified, and the average growth rate for the whole population was estimated to be 6% per day.

5. Similar calculations based on the vertical gradients of nitrate and phosphate indicated what appeared to be relatively lower values for the utilization of these substances. It is pointed out that published data show a normal carbon-nitrogen-phosphorus ratio of 41 : 7.2 : 1 in plankton. The present calculations lead to an analogous estimate of the ratio of carbon production to nitrate utilization to phosphate utilization, but the closest approach to the published ratio that can be attained is 130 : 8.5 : 1. Certain ambiguities in the data are discussed. It remains undecided whether the results indicate actual nutrient deficiency or errors in the assumptions on which the calculation are based.

6. The average wet weight of zooplankton in the upper 400 m of water (or surface to near bottom in coastal water) was 0.84 g per m^3 in the coastal area, about the same in the slope water, 0.15 g in the Gulf Stream, and 0.045 g in the Sargasso Sea. In a deeper stratum of about 400 m to 800 m (uncorrected for wire angle) the quantity of zooplankton varied in different parts of the region from one-fifth to one-half the surface concentration. The averages for total zooplankton were of the order of 30 g wet weight per m^2 of sea surface in the coastal zone and Sargasso Sea, 70 g in the Gulf Stream, and upwards of 80 g in the slope water area.

7. The percentage of organic matter in the zooplankton catches varied widely from a maximum of 16.2 in the coastal collection to 0.7% in the slope water, the low value in the latter being due largely to the predominance of salps. The Sargasso Sea collections averaged 4.8% organic matter. Because of these differences, the regional distribution of zooplankton organic matter was quite different from that of wet weight, with a maximum of 4.8 g per m^2 of sea surface in the coastal water, 1.4 g in the Sargasso Sea, and intermediate values in the Gulf Stream and slope water. Indirect evidence suggests that these regional variations in organic content are largely due to differences in the quantity of stored lipoids (except for a few special cases such as the slope water salps), and this in turn is probably dependent on the concentration of phytoplankton.

8. Experiments on the metabolic rate of Sargasso Sea zooplankton yielded a wide range of results. The average, expressed in terms of daily carbon consumption, was 0.49% of the wet weight and 12% of the zooplankton organic content. A single experiment on coastal water zooplankton gave a value for carbon consumption of 0.6% of the wet weight and 4% of the organic matter per day, and these results are in good agreement with published data on the metabolic rate of *Calanus finmarchicus*.

9. A speculative discussion is presented of zooplankton food prob-

lems. It is estimated that in order to obtain its minimum food requirements, the Sargasso Sea zooplankton population would have to eat somewhere between 7% and 20% of the phytoplankton crop per day, and the coastal water zooplankton would have to eat 6% of the plant population. However, other considerations suggest that the actual consumption in the coastal zone is much greater. It is tentatively concluded that the zooplankton of the coastal zone is better fed than the Sargasso Sea population, which tends to account for its higher organic content and also suggests that it has a higher reproductive rate.

REFERENCES

Bigelow, H. B. and Mary Sears
1939. Studies of the waters of the continental shelf, Cape Cod to Chesapeake Bay. III. A volumetric study of the zooplankton. Mem. Harv. Mus. comp. Zool., *54:* 181–378.

Bogorov, B. G.
1934. Seasonal changes in biomass of *Calanus finmarchicus* in the Plymouth area in 1930. J. Mar. biol. Ass. U. K., N. S. *19:* 585–612.

Clarke, G. L.
1940. Comparative richness of zoöplankton in coastal and offshore areas of the Atlantic. Biol. Bull. Woods Hole, *78:* 226–255.

Clarke, G. L. and D. D. Bonnet
1939. The influence of temperature on the survival, growth and respiration of *Calanus finmarchicus.* Biol. Bull. Woods Hole, *76:* 371–383.

Clarke, G. L. and D. F. Bumpus
1940. The plankton sampler—an instrument for quantitive plankton investigations. Limnolog. Soc. Amer., Spec. Pub. No. 5: 1–8.

Fleming, R. H.
1940. The composition of plankton and units for reporting populations and production. Proc. 6th Pacif. Sci. Congr., Calif. (1939), *3:* 535–540.

Harvey, H. W.
1934. Measurement of phytoplankton population. J. Mar. biol. Ass. U. K., N. S. *19:* 761–773.
1937. Note on selective feeding by *Calanus.* J. Mar. biol. Ass. U. K., N. S. *22:* 97–100.

Jesperson, P.
1935. Quantitative investigations on the distribution of macroplankton in different oceanic regions. Rep. "Dana"-Exped., *2* (7): 1–44.

Leavitt, B. B.
1935. A quantitative study of the vertical distribution of the larger zoöplankton in deep water. Biol. Bull. Woods Hole, *68:* 115–130.
1938. The quantitative vertical distribution of macrozoöplankton in the Atlantic Ocean basin. Biol. Bull. Woods Hole, *74:* 376–394.

Lohmann, H.
1920. Die Bevölkerung des Ozeans mit Plankton nach den Ergebnissen der Zentrifugenfänge während der Ausreise der "Deutschland" 1911. Zugleich ein Beitrag zur Biologie des Atlantischen Ozeans. Arch. Biontol. Berl., *4* (3)*:* 1–617.

Marshall, S. M., A. G. Nicholls and A. P. Orr
1934. On the biology of *Calanus finmarchicus.* V. Seasonal distribution, size, weight and chemical composition in Loch Striven in 1933, and their relation to the phytoplankton. J. Mar. biol. Ass. U. K., N. S. *19:* 793–827.
1935. On the biology of *Calanus finmarchicus.* Part VI. Oxygen consumption in relation to environmental conditions. J. Mar. biol. Ass. U. K., N. S. *20:* 1–27.

Nakai, Z.
1942. The chemical composition, volume, weight and size of the important marine plankton. J. oceanogr. Soc. Japan, *1:* 45–55, 146 (Eng. abstr.).

Riley, G. A.
1939. Plankton studies. II. The western North Atlantic, May–June, 1939. J. Mar. Res., *2:* 145–162.
1941. Plankton studies. IV. Georges Bank. Bull. Bingham oceanogr. Coll., *7* (4)*:* 1–73.

Seiwell, H. R.
1935. The cycle of phosphorus in the western basin of the North Atlantic. I. Phosphate phosphorus. Pap. phys. Oceanogr. Meteorol., *3* (4)*:* 1–56.

Sverdrup, H. U.
1938. On the explanation of the oxygen minima and maxima in the oceans. J. Cons. int. Explor. Mer, *13:* 163–172.

Zeuthen, Erik
1947. Body size and metabolic rate in the animal kingdom. C. R. Lab. Carlsberg (Ser. Chim.) *26* (3)*:* 17–161.

APPENDIX

STATION* AND MISCELLANEOUS SURFACE ANALYSES. TEMPERATURE IN °C, PHOSPHATE AS MG-AT OF P PER M³, NITRATE AND NITRITE AS MG-AT N PER M³, PLANT PIGMENTS AS HARVEY UNITS PER M³

Station Date N Lat. W Long.	*Depth (meters)*	*Temp. (°C)*	*P*	*NO_3*	*NO_2*	*Plant pigments*
4603	1	23.8	—	1.7	.09	—
17 VII	100	14.0	—	6.3	.12	—
39–12	200	11.9	—	1.0	.02	—
70–29	300	9.1	—	—	.00	—
	400	7.1	—	6.5	.05	—
4604	1	26.2	—	.4	.00	—
20 VII	25	25.7	—	.7	.00	—
34–56	50	22.8	—	.6	—	—
65–34	100	22.1	—	1.3	.08	—
	150	19.1	—	2.2	.00	—
	200	18.8	—	2.2	.23	—
	300	18.4	—	2.4	.00	—
	400	17.8	—	4.1	.00	—
4605	1	27.7	—	.9	.06	—
25 VII	100	18.9	—	3.1	.02	—
33–30	200	18.9	—	2.3	.08	—
62–48	300	18.3	—	2.7	.05	—
	400	17.8	—	6.0	.15	—
4606	1	27.6	—	1.1	.07	—
26 VII	25	24.1	—	3.7	.07	—
35–04	50	21.3	—	.7	.10	—
60–36	100	19.2	—	2.3	.23	—
	150	18.8	—	2.5	.16	—
	200	18.4	—	4.6	.09	—
	300	18.1	—	2.6	.10	—
	400	17.5	—	5.2	.07	—
4607	1	26.3	.03	.0	.00	280
30 VII	100	20.3	.12	.0	.07	360
37–15	200	18.5	.10	1.9	.04	300
50–21	300	18.2	.11	2.3	.00	170
	400	17.8	.17	2.0	.00	ca .0
4608	1	25.0	.06	.8	.28	150
1 VIII	100	17.7	.10	2.6	.10	110
34–52	200	16.4	.31	4.7	.05	50
43–29	300	15.3	.37	3.7	.07	ca .0
	400	14.1	.47	6.4	.06	ca .0
4609	1	25.2	.07	.7	.17	70
3 VIII	100	17.8	.12	4.7	.15	ca .0
32–29	200	16.7	.21	3.2	.09	ca .0
41–00	300	15.6	.29	4.8	.08	ca .0
	400	15.1	.27	.7	.08	ca .0

* Tables include only the observations considered pertinent to the plankton problem. Precise temperature readings, oxygen concentrations and deep water distributions will be published subsequently.

Station Date N Lat. W Long.	*Depth (meters)*	*Temp. (°C)*	*P*	*NO_3*	*NO_2*	*Plant pigments*
4610	1	25.6	.00	.5	.15	—
7 VIII	25	25.4	.05	.2	.08	—
30–04	50	22.7	.02	.1	.10	—
42–02	100	19.8	.16	.0	.18	—
	150	18.7	.10	.7	.22	—
	200	17.8	.18	2.8	.18	—
	300	17.2	.25	4.2	.24	—
	400	16.0	.30	5.3	.12	—
4611	1	25.8	.10	.9	.10	190
13 VIII	100	17.7	.00	2.6	.08	210
32–22	200	15.9	.19	3.0	.08	160
38–50	300	14.9	.32	4.8	.07	—
	400	13.9	.44	7.0	.05	—
4612	1	25.4	.08	.0	.35	200
15 VIII	100	18.4	.01	1.0	.31	420
34–21	200	16.9	.23	3.8	.22	110
33–29	300	15.4	.31	4.7	.23	110
	400	14.5	.37	5.8	.31	—
4613	1	23.4	.04	.6	.29	—
21 VIII	25	22.8	.05	.3	.32	—
35–42	50	18.7	.01	.2	.21	—
27–47	100	16.0	.11	1.2	.39	—
	150	15.4	.21	4.0	.21	—
	200	14.7	.38	2.8	.26	—
	300	13.5	.51	7.3	.25	—
	400	12.5	.61	7.0	.11	—
4614	1	25.6	.06	.0	.20	—
23 VIII	100	19.2	.00	.7	.17	—
33–34	200	17.3	.17	4.1	.27	—
30–46	300	16.0	.25	6.5	.21	—
	400	15.0	.37	10.5	.20	—
4615	1	28.2	.05	.0	.18	—
31 VIII	25	27.0	.03	.0	.22	—
29–00	50	—	.03	.7	.16	—
48–31	100	21.8	.09	.8	.21	—
	150	—	.09	.0	.44	—
	200	18.4	.16	.9	.38	—
	300	—	.26	2.0	.22	—
	400	16.5	.35	5.7	.28	—
4616	1	29.0	.24?	.2	.25	—
2 IX	25	28.2	.00	1.0	.24	—
29–03	50	—	.04	.0	.28	—
55–18	100	20.6	.05	.0	.27	—
	150	—	.06	.9	.28	—
	200	18.3	.18	2.2	—	—
	300	—	.24	2.7	.22	—
	400	16.8	.28	3.1	.21	—

Station Date N Lat. W Long.	*Depth (meters)*	*Temp. (°C)*	*P*	*NO_3*	*NO_2*	*Plant pigments*
4617	1	29.2	.05	.0	.54	—
4 IX	25	28.8	.03	.7	.29	—
30–22	50	—	.00	1.1	.33	—
60–31	100	20.4	.00	1.3	.29	—
	150	—	.03	1.2	.35	—
	200	18.5	.04	1.7	.28	—
	300	—	.13	2.7	.24	—
	400	17.6	.20	4.0	.25	—
4618	1	28.5	.02	.0	.30	—
12 IX	25	28.5	.00	.0	.30	—
38–21	50	—	.00	1.0	.27	—
69–53	100	22.1	.01	.1	.26	—
	150	—	.03	.2	.29	—
	200	19.1	.05	.4	.33	—
	300	—	.15	1.9	.23	—
	400	17.7	.19	3.2	.20	—

Oxygen Concentrations (ml per l) Used to Compute Net Production

Station	*25 m*	*50 m*	*100 m*
4606	4.99	5.15	4.68
4610	4.62	5.36	5.05
4615	4.57	5.01	4.73
4616	4.69	5.35	5.10
4617	4.45	5.13	5.00

Surface Water Analyses

Date	*N. Lat.*	*W. Long.*	*Temp.*	*P*	*NO_3*	*NO_2*	*Plant pigments*
6 VIII	41–17	70–59	20.8	—	—	—	6,000
6 VIII	40–55	70–58	22.6	—	—	—	4,000
8 VIII	40–21	70–53	23.3	—	—	—	18,000
7 VIII	39–20	70–42	24.5	—	—	—	11,000
7 VIII	28–55	41–19	25.2	.12	.0	.14	190
9 VIII	32–00	42–26	25.6	.19	.0	.17	110
22 VIII	34–33	29–18	—	.03	.3	.00	110
24 VIII	33–07	32–29	25.6	.00	.0	.17	90
26 VIII	30–49	36–25	—	.08	.0	.23	100
30 VIII	29–00	44–44	—	.10	.0	.33	110
1 IX	29–03	51–34	28.2	.00	.0	.20	80
2 IX	29–02	54–50	—	.12	.0	.25	110
4 IX	30–19	60–07	29.0	.01	.0	.22	150
10 IX	34–17	65–55	29.2	.04	.0	.42	110

NOTE TO AUTHORS

The pages of this Journal are open to all contributors, and manuscripts may be submitted to any member of the Board of Editors or to the Managing Editor. Publication will be made only in the English language, but copy may be submitted for consideration in other languages. Aid in translation of accepted manuscripts will be given when possible.

All manuscripts should be typewritten with double space, carefully corrected, with footnotes continuously numbered and set off in the body of the text. If an article is to appear as a contribution from another institution, it should be approved by that institution before submission; contribution numbers of such institutions are submitted with the return of galley proofs.

Each author receives 50 reprints (with covers) free of charge up to a maximum of 150 free reprints of any single article in the case of joint authorship by three or more authors. Additional reprints with or without covers may be ordered by authors or institutions at the cost of printing. Orders for reprints are given when galley proofs are returned.

Both galley and page proofs are submitted to the senior author of each article.

Tables should be prepared for clarity and as near as practicable to Journal pattern, without ruled lines if possible. Table footnotes should appear at the base of each Table.

Figures are required for consideration simultaneously with the manuscript. Original figures of inconvenient size for forwarding may be represented by photographs on first submission of a manuscript; originals are required for line cuts. Illustrations requiring half-tone reproduction should be avoided as far as possible, but they will be accepted when necessary. *It is very important that the proper size of characters and symbols should be used for the lettering of illustrations.* A diagram for determining suitable heights of letters for original drawings that will be reduced, together with an explanatory note, has been published in this Journal, Vol. III, No. 1: 89–91; reprints will be submitted to authors on request. Mechanical lettering is recommended. If facilities for lettering are not available to the author, it is preferable to receive illustrations with lettering indicated in pencil only, to be finished in ink in the editorial office. All illustrations will be subject to at least a minimum reduction at the rate of 1 : 1.2. Drawings which have been heavily patched and pasted must *not be rolled* for mailing. The legends for the figures, identified by corresponding figure numbers on the illustrations also, should be typed in sequence at the end of the manuscript.

A complete and *accurate list of references* in alphabetical order at the end of each article is essential. References may be indicated in the text by author's name (or names) and *year of publication*, or by italicized numbers which correspond to the numbers given in sequence in the list of references. Each reference should include the last name followed by at least two initials or first name; co-authors should be listed with the initials or first name preceding the last name. The date should be indented on the line below, followed by a complete title *verbatim* in lower case except for the first word. Publisher, city and pagination should be indicated for books; for periodicals (abbreviations from "World List of Scientific Periodicals," Oxford, as far as possible), the volume (italicized), number and pagination should be given.

Additional information regarding the preparation of manuscripts will be furnished by the Managing Editor.

TABLE OF CONTENTS

Page

1. **Yasuo Miyake and Masami Koizumi:** The measurement of the viscosity coefficient of sea water 63

2. **Yasuo Miyake and Tomosaburo Abe:** A study on the foaming of sea water 67

3. **Robert O. Reid:** The equatorial currents of the eastern Pacific as maintained by the stress of the wind 74

4. **Gordon A. Riley and Samy Gorgy:** Quantitative studies of summer plankton populations of the western North Atlantic 100